Winds Over Wildlands—

A Guide for Forest Management

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Wildland area management and protection are affected in many ways by the behavior of the wind. Wind distributes the seed on which natural regeneration of forest trees and range plants depends. Timber cutting practices can be varied to take advantage of this, to red eth hazards of windthrow, or to influence snow accumulation, depending on knowledge of local winds. Patterns of damage from air pollutants are determined by the wind. Forest diseases are spread by airborne spores. Forest fires are especially sensitive to wind behavior.

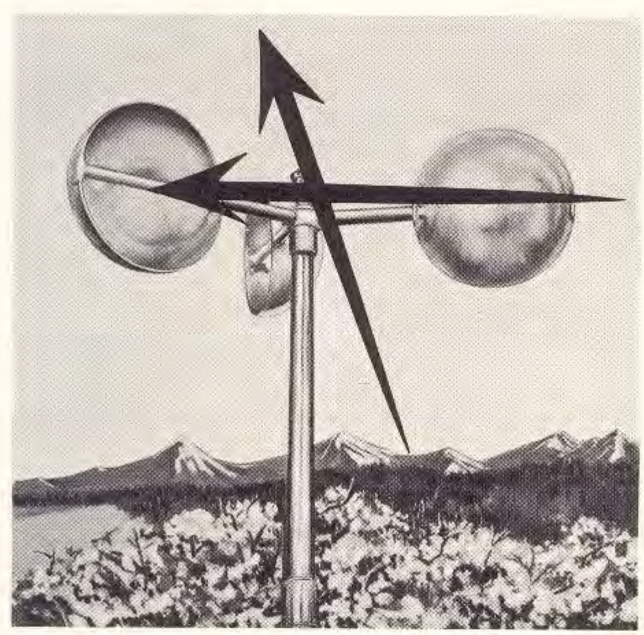
Hence, practical management often requires tat the land manager or technician evaluate the winds concerned in special ways to meet his peculiar needs. For example, he may wish to know only the wind speed and direction as measured by conventional surface wind instruments. Other problems are concerned also with the windflow aloft, perhaps to heights of several thousand feet. Still others are related to vertical air motion in the form of updrafts and downdrafts at various heights or to various combinations of flow in both vertical and horizontal directions. In all cases, however, the detailed structure of the airflow that must be evaluated is above a local problem area. Such an area may have a perimeter encompassing a single tree in a timber stand or extend or several miles of mountain river canyon.

Interpretation of wind for these wildland applications involves a concept of wind beyond that usually implied in common meteorological use. Wind is defined as air in motion with respect to the earth's surface. Commonly, thought, wind is understood to refer only to the component of motion in a horizontal direction since this is the dominant motion in the atmosphere. Air moves thousands of miles around the hemisphere. Vertical motion is negligible relative to the distance around the hemisphere and is limited to the lowest layer of the atmosphere, called the troposphere, roughly 5 to 8 miles deep over most of North America. This vertical distance of motion is quite significant, however, compared with the relatively short horizontal distances involved in most local wildland problems. Hence, for many applications, wind must be interpreted in terms of actual air speed and its true direction whether in horizontal flow, up, down, or in rotational or spiral flow as in a whirlwind.

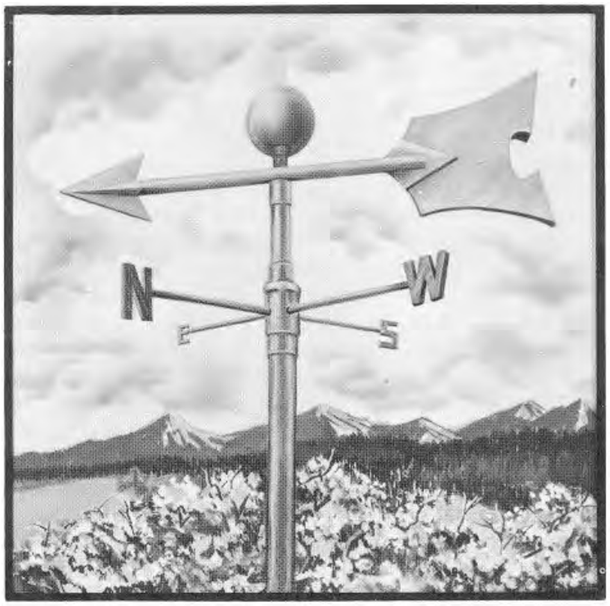
Conventionally, winds are separated into surface winds and winds aloft. There is no sharp line between them, but rather a blending of one into the other. Surface winds are considered as winds measured with instruments mounted on surface-borne masts or towers. Winds aloft are those to an indefinite height and measured with airborne equipment.

Surface wind speed and direction are measured by anemometers and wind vanes. These are usually of conventional design and, in line with the usually accepted definition of wind, indicate only the horizontal components of air motion. Cup anemometers respond to air blowing up or down through the rotating cup assembly, but their records are interpreted as horizontal motion. Wind vanes only point in the direction from which the wind blows. Protection organizations in the United States have accepted 20 feet above open level ground as the standard exposure for wildland surface wind sampling.

Windspeeds are measured and reported in either miles per hour or knots. Miles per hour are commonly used in referring to surface windspeeds on land for civilian use, while knots are the more common measure for aeronautical and marine applications.



Cup anemometer.



Wind vane.

A knot is 1 nautical mile per hour, equal to 1.15 land or statute miles per hour.

Surface observing and reporting stations are usually a few to many miles apart. Portable or temporary instrument installations are often useful locally, but even with these it is usually impossible to sample sufficiently to obtain a significant pattern of air motion or its changes with time over a sizable area. Thus, the principal value of surface instrument measurements is often their indication only of the general intensity of movement applicable to a small topographic unit.

Winds aloft are determined most commonly by tracking helium-filled balloons from the surface up through the atmosphere. The simplest system employs a pilot balloon tracked visually with a theodolite. Assuming constant rate of rise of the balloon, frequent readings of elevation and azimuth with the theodolite allow computation of the wind speed and direction. Errors are introduced when the air is such that the balloon ascent is not at a constant rate. Adding to the balloon a radiosonde unit which transmits temperature, moisture, and pressure data during ascent reduces these errors somewhat. The most refined of present systems has the further addition of a self-tracking radar that measures elevation, azimuth, and slant distance of the balloon from the observing station. This unit, known as a rawinsonde, gives quite accurate upper air information. All of these measurements are referred to as soundings, meaning upper air observations.

The speed and direction of winds aloft are sampled at regular hours each day at selected weather stations scattered over the continent. These stations are frequently 100 miles or more apart. Although winds aloft tend to be somewhat more uniform than winds near the surface, exceptions are frequent. Thus, the wind structure over an area some distance from a sampling station may differ considerably from that indicated by the sounding.

Of all wildland applications forest fire suppression should be based on probably the most exact interpretation of local winds; yet such interpretation is at best an estimate. Air motion in the atmosphere is extremely complex, and available instruments and sampling techniques give only rough approximations of actual air behavior. Operationally the interpreter of local weather uses both surface and upper air measurements as indicators of the kind of weather system in which he is working. From there on, what his eye can see and his skin feel, supplemented by his knowledge of wind behavior in general, are his principal tools. Ripples on open water, foliage, dust, fire and smoke, haze, clouds, soaring birds, temperature, and fell of the wind on the body are some useful indicators. How to use them is an art developed through long personal experience. This book is designed to shorten the time of familiarization by presenting basic principles of air motion and common airflow patterns encountered in wildland problems.

Principles of Air Motion-Properties of Air

Air is a mixture of gases, mostly nitrogen and oxygen. And even though not heavy in comparison with other familiar substances, it does have measurable mass and responds accordingly to the force of gravity. Thus, a 1-square-inch column of air the height of the atmosphere weighs 14.7 pounds at sea level.

Within the air this weight is equivalent to the more familiar 14.7 pounds per square inch pressure at sea level. Normal sea-level pressure is equal to that exerted by a mercury column 29.92 inches tall. This is equivalent to 1,013 millibars (mb.), another common pressure measurement used in meteorology.

Pressure observations made at weather stations are called surface pressure or station pressure observations. These are reduced to sea-level pressure by standard corrections for station altitude.

Air at the outer limits of the atmosphere is extremely rare, its pressure approaches zero, and each cubic foot containing only a few molecules weighs virtually nothing. However a cubic foot at sea level—compressed by all those cubic feet above and containing many more molecules—weighs about 0.08 pound at 32°F. It follows that both air density and atmospheric pressure decrease with increasing altitude. The 500-mb. level at mid-latitudes is at an average altitude of about 18,000 feet. Thus, nearly half the weight of the atmosphere is below this height.

Air that is compressed is warmed. Expansion reverses the process and cools the air. Thus, rising air cools and settling air warms. The converse is also true; air that is warmed or cooled by other mechanisms expands or contracts. Thus, the 0.08-pound cubic foot of air at sea level warmed to 60°F. weighs only 0.075 pound and at 90°F. only about 0.072 pound. Air density therefore changes with both temperature change at constant pressure and change in pressure.

Air can be diluted by water vapor in the atmosphere. The amount of water vapor that can be held in the atmosphere is determined only by the temperature of the atmosphere. The maximum is about 5 percent by volume near the surface in warm regions, decreasing aloft with the lower temperatures of the higher altitudes. The temperature at which saturation is reached and when condensation may begin, as cloud water droplets for example, is called the temperature of the dewpoint. Condensation occurs at the dewpoint if there are suitable condensation nuclei or other surfaces present. This is the usual case, although the absence of nuclei may occasionally require lower temperatures for condensation.

Water vapor molecules do not weigh as much as air molecules. Thus, a mixture of dry air and water vapor, or a moist atmosphere, is less dense than a dry atmosphere at the same temperature and pressure.

Air moves in the free atmosphere in response to either of two principal forces generated within the atmosphere. The first of these is buoyant force whereby warm, less dense air is forced upward by surrounding cooler and more dense air, and is caused by changes in air density due to heating and cooling. The amount of upward thrust depends on the temperature difference. It results in vertical circulation, commonly known as convection or, more properly, free convection.

The second of these forces is horizontal pressure gradient which causes air to move horizontally. The flow is from high to low pressure. Convection is a primary source of these pressure differences. Expanding and rising air flows outward at the top of the uplift into surrounding areas. Over extensive heating surfaces this outflow may take place near the top of the troposphere. Over lesser areas, it may occur at various lower altitudes. In either, the result is lowering of pressure over the warmer areas and increasing pressure over the cooler areas.

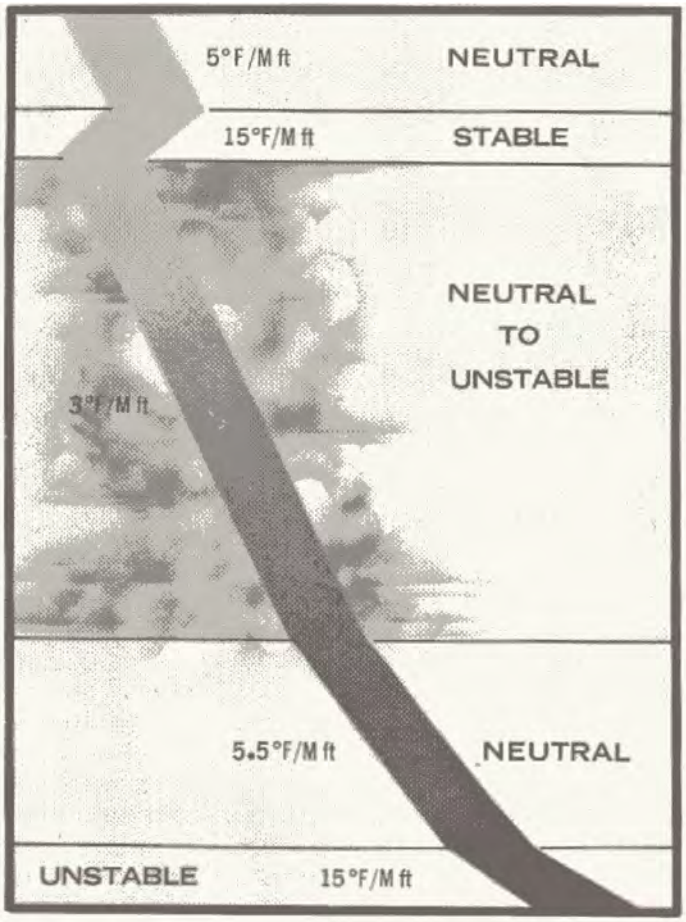
The energy required to generate these forces originates primarily through contact of the air with the earth's surface. Irregular heating and cooling of the surface result in temperature differences in the overlying air. These produce both buoyant forces and horizontal pressure differences. Over warm moist surfaces, only part of this thermal energy is converted immediately into air expansion. Tremendous quantities of energy are expended in evaporating water into the atmosphere. But the resulting water vapor upon mixing and flowing with the air carries this energy with it. Upon condensation as cloud droplets or precipitation, the same amount of heat is released to warm the air that was used in the evaporation. This warming may either initiate further motion or increase airspeeds in regions far removed from where the original evaporation took place.

Atmospheric Stability

Conditions in the atmosphere that are adverse o or favorable toward vertical circulation are usually referred to as stable and unstable, respectively. A stable atmosphere resists or damps vertical motion; an unstable one encourages it. Stability and instability are defined primarily by the temperature change with height above the surface, or the temperature lapse rate.

If the troposphere were thoroughly mixed, and dry, the temperature would decrease regularly at 5.5°F. per 1,000 feet of elevation. This is called the dry adiabatic lapse rate and is the rate at which a sample of dry air cools when lifted. The troposphere would then be neutral. Mixing in great depth is never complete, however, so various portions of the atmosphere aloft often differ in stability.

Stability is characteristic of a segment of unsaturated atmosphere in which the temperature decrease is less than 5.5°F per 1,000 feet of altitude. Maximum stability occurs in the event of a



The degree of atmospheric stability at any height depends on the temperature lapse rate.

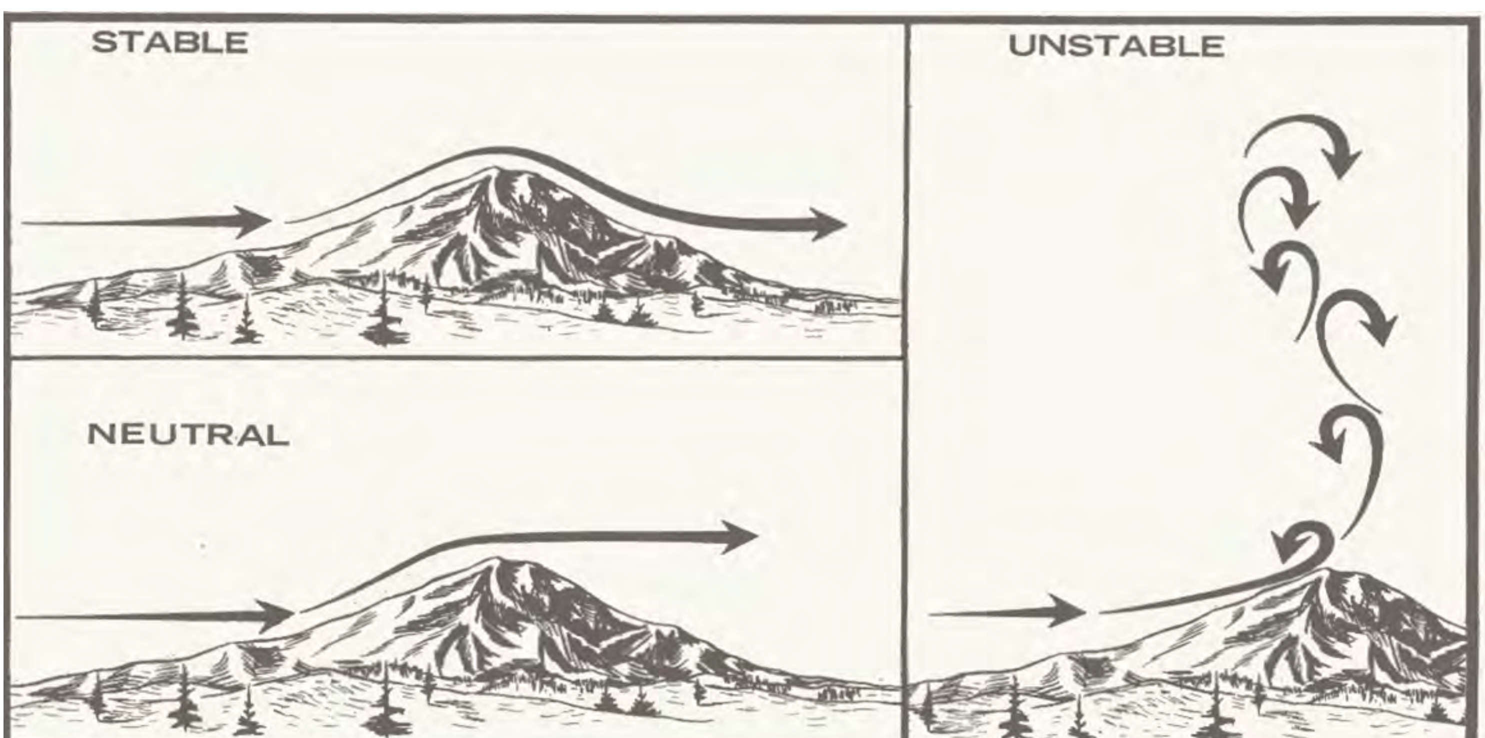
temperature inversion, a condition in which there is an increase in temperature with height.

Air is unstable when the temperature decrease with height is greater than the adiabatic lapse rate. Lapse rates greater than adiabatic are known as superadiabatic lapse rates. Moisture in rising air tends to increase the instability. Moist air, less dense than dry air, rises, and cools at the dry adiabatic lapse rate until it reaches the dewpoint temperature where condensation begins. Cooling above this level continues, but at a slower rate (the moist adiabatic lapse rate) because of the heat liberated by condensation. The moist adiabatic lapse rate varies with temperature, but is usually in the order of 2° to 3°F. per 1,000 feet.

Air forced upward in a stable atmosphere cools at the adiabatic rate and thus becomes colder than its surroundings. This causes it to settle back to its original level. Air forced downward becomes warmer than its surroundings and rises again.

Air in any part of a neutral atmosphere when physically lifted or lowered would change temperature at the same rate as its surroundings. It would thus remain in equilibrium at any level to which moved and immediately come to rest upon removal of the lifting or lowering force.

Vertical motion in an unstable atmosphere, once initiated, is accelerated. Cooling only

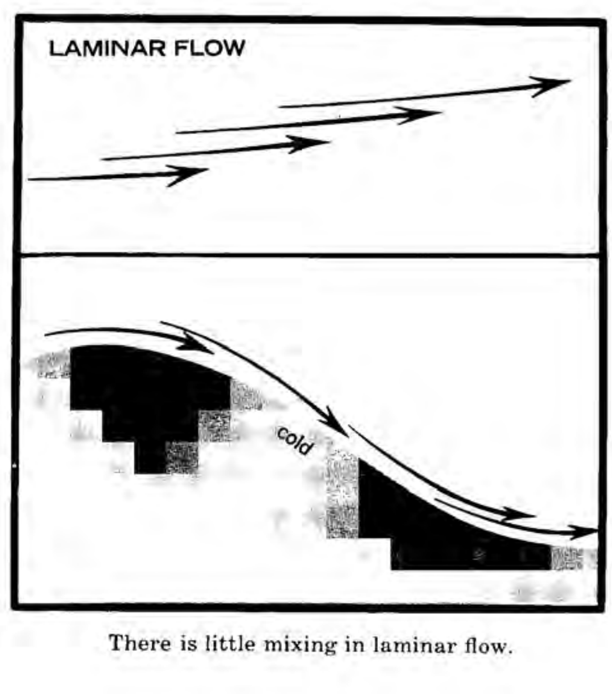


Air forced upward drops down, stays up, or continues to rise, depending on local condition of the atmosphere.

at the adiabatic rate, the rising air becomes increasingly warmer than its surroundings with corresponding increases in buoyant force and speed of rise. Condensation, as in cumulus cloud formation, is a frequent contributor to strong instability and high rates of ascent within the cloud cell. The cloud bases are at or near the saturation level.
Air near the surface becomes stable with surface cooling as at night, and very stable if the cooling is sufficient to create a surface inversion. In many climates these are a nightly occurrence during much of the year. Inversions also occur aloft at any elevation when warm air overrides a layer of cool air or cool air slides under warmer air. The elevation of maximum temperature in an inversion is known as the top of the inversion. The inversion, reflecting a highly stable situation, strongly resists any vertical motion or transfer of energy through it.
Instability is frequent near the ground on clear sunny days. Air near the surface, at least, is unstable if the ground feels warm to the touch, and is markedly unstable if the ground feels hot. Instability favors convection, a necessary mechanism for mixing in the atmosphere. The initial lifting may be initiated by the air flowing over rising topography or by local disturbances in the air near the ground. Rise starts spontaneously when the lapse rate reaches a critical value. In quiet uniform air near the ground, this is in the neighborhood of 18°F. per 1,000 feet. Under some atmospheric conditions in a natural environment, it can probably be greater.
The height to which convective activity extends depends on the rate of surface heating, on water vapor content, on the efficiency with which mixing takes place aloft, and on the presence or absence of barriers to vertical circulation. The amount of water vapor determines how much heat will be available from condensation. Mixing with surrounding air tends to lower the height of vertical penetration. Convection in otherwise quiet air is not very effective in mixing, but in the presence of horizontal windspeeds it contributes significantly to mixing and attainment of adiabatic lapse rates. Wind shear aloft between air layers differing in wind speed and direction generates turbulence and may thus assist in the mixing. Inversions at any height in the atmosphere inhibit circulation through them. Wind shear and inversions often occur together.
Cumulus clouds are often the best indicators of local instability with poor mixing aloft. The rising air in a growing cloud tends to retain its own temperature and moisture characteristics until the cloud breaks up. It may then gradually mix with the surrounding air.
Upward convection in unstable air is accompanied by downward settling air at a generally slower rate over surrounding areas. This is true both in the large-scale circulations around high and low pressure systems and in situation of local instability. In stable air situations, however, downward-flowing air may slide under less dense air and slowly lift it. In such situations it is the downward flow that is the more significant. Typical of these are surface winds of the foehn type and those resulting from nighttime cooling and from thunderstorm downdrafts.

Types of Airflow

Air moving in response to pressure differences, but undisturbed by surface friction or vertical convection, flows smoothly in streamline or laminar fashion. Laminar flow is a suggestive term indicating air moving along in flat sheets with each successive thin layer sliding over the next. Laminar or near-laminar flow is typical of stable air above the surface moving at low speeds. In surface winds it is characteristic of flow below a nighttime inversion rather than of daytime winds. Vertical mixing is negligible.
When either friction or heating at the ground surface occurs, surface winds become turbulent. Air particles no longer follow straight lines, but move at varying speeds and constantly changing directions.



There is little mixing in laminar flow.

Either roughness or heating of the surface causes turbulent mixing.

At the surface, turbulence is commonly identified in terms of eddies, whirls, and gusts; aloft it is more frequently associated with "bumpy" flying.

Surface friction produces mechanical turbulence in the air. The flow of stable air near the surface is similar to the flow of water in a creek bed. Currents in stable air at low speed tend to follow the general contours of the landscape. But when the speed increases—as when a creed floods—the current "tumbles" over and around hills and ridges, structures, trees, and other obstacles and sets up eddies in all directions. Mechanical turbulence increases with both windspeed and roughness of the surface.

Thermal turbulence is associated with instability and convective activity. It is similar to mechanical turbulence in effects on surface winds, but extends higher in the air aloft. In flat country it is also less definitely tied to any one spot. Since it is the result of surface heating, thermal turbulence increases with the intensity of surface heating and the degree of instability indicated by the temperature lapse rate. Mechanical and thermal turbulence frequently occur together, each magnifying the effects of the other.

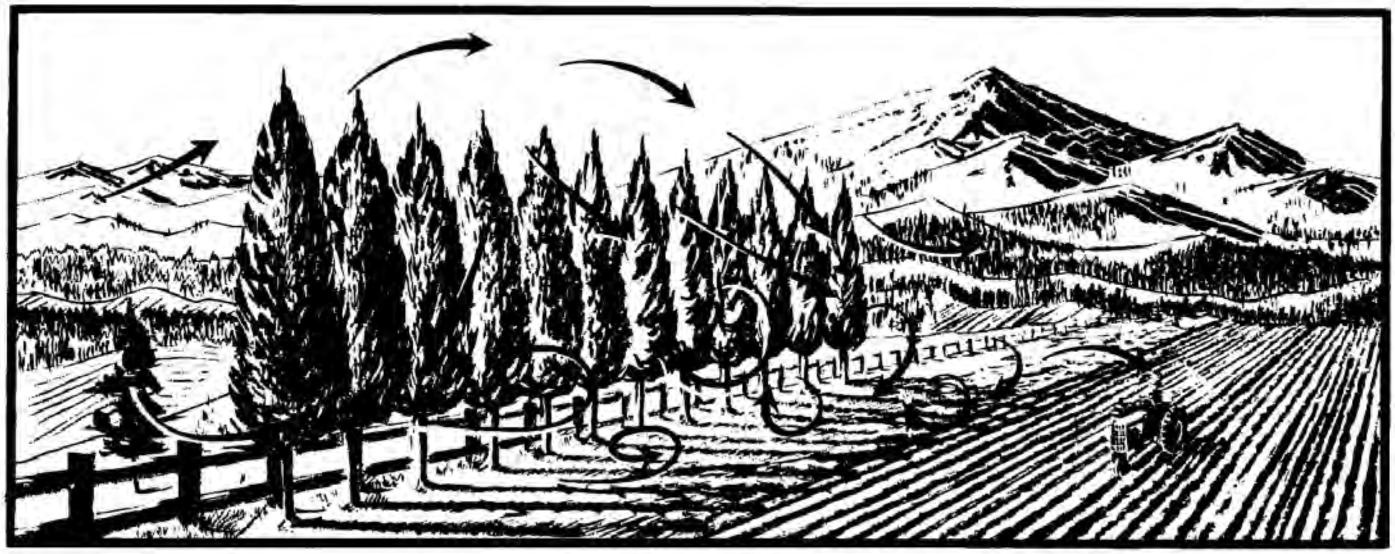
Thermal turbulence induced by the combination of convection and horizontal wind is the principal mechanism causing energy interchange between the surface and the winds aloft. Unstable air warmed at the surface rises to mix and flow along with the winds above. This turbulent flow also brings any higher windspeeds aloft down to the surface, usually in spurts and gusts.

Eddy formation is a common characteristic of turbulent flow. Every solid object in the wind path creates eddies on its lee side. Their sizes, shapes, and motions are determined by the shape of the obstacle and the speed and direction of the wind. Besides these, the general roughness of the area contributes to formation of larger eddies that move over the landscape. Eddy motion is also characteristic of thermal turbulence.

Although eddies may form in the atmosphere with their axes of rotation in virtually any plane, it is usual to distinguish only those that have predominantly vertical or horizontal axes. A whirlwind or dust devil is a vertical eddy. Large, roughly cylindrical eddies that roll along the surface like tumbleweeds are horizontal eddies.

Eddies associated with individual fixed obstructions tend to hold a more or less stationary position in the lee of the obstruction, although secondary eddies frequently break off and move on downwind. For most obstructions the general rule of thumb is that an obstacle affects the windstream for a distance downwind 8 to 10 times the height of the obstacle over which the wind flows.

Rotation speeds in eddies are often much greater than the average windspeeds measure with mechanical anemometers. These higher speeds are often of short duration at any point, except where stationary eddies are formed, but are still significant for many purposes. Whirlwinds, for example, develop speeds capable of lifting sizable objects. Eddies moving with the general wind account for the principal short-term changes in wind speed and direction known as gustiness.



Obstructions to airflow produce eddies.

General Circulation

Differences in heating of the troposphere above warm and cool regions set up pressure systems that dominate the airflow over the North American continent. The equatorial regions receive far more solar energy than they lose, while over the polar regions there is a net loss. In between, there are many regions throughout the hemisphere where temperature contrasts influence pressure and wind on a lesser scale. Typical of these are land and sea surfaces. Water surfaces both heat and cool more slowly than land. Thus the oceans may be cool in relation to land in summer, and the reverse in winter.

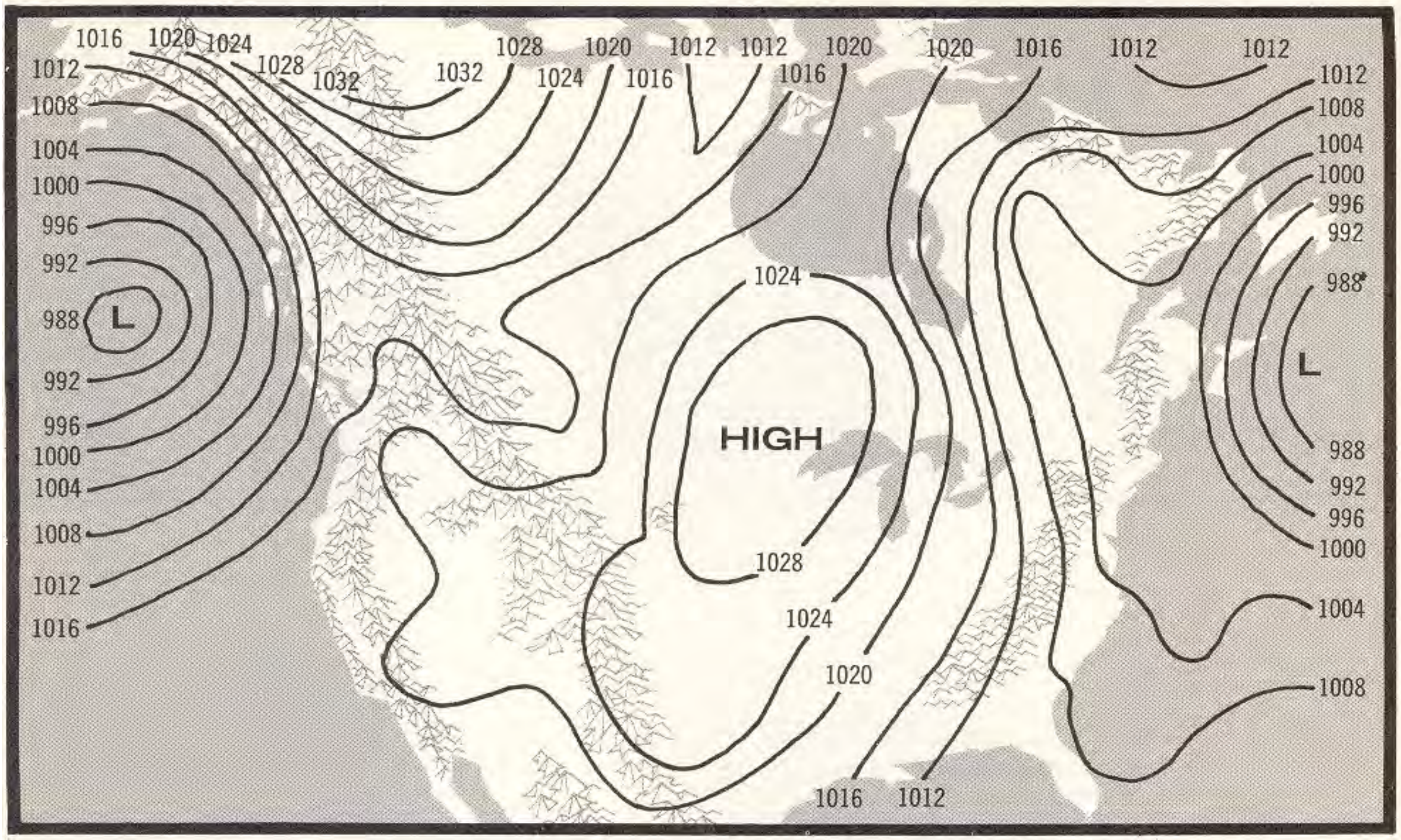
Air that remains very long over any region acquires the moisture and temperature characteristics of the region. Air acquiring these regional characteristics becomes identified as an air mass. Air masses are continually building up, migrating, and eventually decaying or acquiring other characteristics in other regions.

The windflow patterns associated with air masses and their pressure systems are always from high pressure to low pressure, but are extremely complex. The leading edge of an air mass is a front. The effects of fronts on winds near the surface are discussed later, but in the general circulation there are other considerations. The earth's rotation, for example, prevents airflow in a linear direction from high to low pressure over any great distance.

In the Northern Hemisphere, the rotation bends the airflow to the fight.

The major pressure systems governing winds over the continent are shown on *weather maps* compiled from simultaneous observations over the continent. It takes several of these maps plotted for different altitudes to obtain the complete picture. The surface weather map shows the sea-level pressure distribution in the form of isobars, lines of equal pressure. The isobars are indicators of wind direction and relative speed expected at the top of the friction layer, the region in which surface friction and thermal turbulence occur.

The pressure gradients are always straight across the isobars, but bending of the wind to the right causes general circulation above the mixing layer to be clockwise in high pressure systems and counterclockwise around centers of low pressure. Above the friction or mixing layer and on up through the troposphere, wind blows nearly parallel to the isobars instead of across them. Spacing of the isobars indicates steepness of the pressure gradients and, thereby, relative windspeeds. Close spacing indicates generally high speeds, and wide spacing relatively low speeds. Knowledge of the large-scale circulation at any time is helpful in interpreting wind for field use, but can only be obtained at central weather offices.



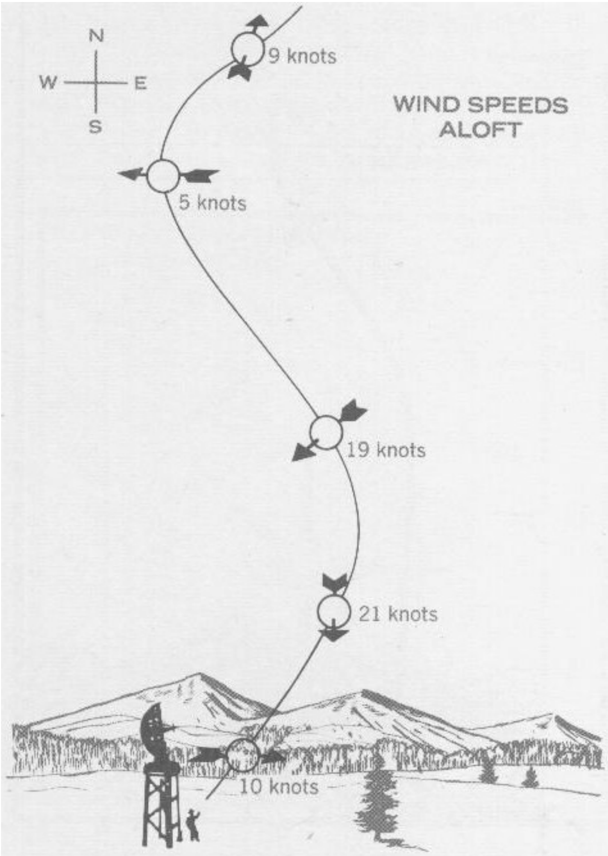
Sea-level pressures are shown by isobars on the surface weather map.

Local Winds Aloft

Locally, the wins aloft may deviate somewhat from those indicated by the large-scale circulation pattern. Although more steady on the average than surface winds, the winds aloft do change as the pressure centers move and as large-scale pressure effects are modified by other factors. Secondary and more transient HIGHS and LOWS, for example, influence local winds aloft, although they are often too small to show on weather maps.

The lowest layer of the winds aloft, the mixing layer is a transition zone. Through it, the winds at the top of the layer are modified to produce the general winds at the surface. These modifications are discussed in the next section, "General Winds Near the Surface."

Pressure systems high in the troposphere may differ from those near the surface. At progressively higher altitudes closed pressure systems (isobars in concentric ovals or circles) are fewer, and fronts are more difficult to identify or locate. Furthermore, it is common for the troposphere to be stratified or layered, and there may be gradual changes in the distribution of HIGHS and LOWS with height. These changes produce different wind speeds and directions in the separate layers. With strong stratification, wind direction often changes abruptly from one layer to the next. The difference in direction may be anything from a few degrees to complete reversal. In the absence of marked stratification above the friction level, wind direction at all levels tends to be more uniform, even though the speed may change with elevation. A common source of stratification in the air aloft is the overriding or underrunning of one air mass by another. Thus, the layers are often unlike in temperature, moisture, motion, or in any combination of the three. Marked changes in either wind speed or direction between atmospheric layers often signify an inversion which damps or cuts off vertical circulation, whether convection over a fire or natural circulation as in cumulus cloud formation. Even though a windspeed profile of the upper air might



Winds aloft profile in a stratified atmosphere.

Indicate only nominal airspeeds, the relative speeds of two air currents flowing in opposite directions may produce visible wind shear effects. Clouds at different levels moving in different directions, tops being blown off growing cumulus clouds, and rising smoke columns that break off sharply and change direction are common indicators of disrupted vertical circulation patterns.

Local wind-aloft profiles commonly fall into one or another of several general types. The four types illustrated by soundings on different dates at one station and reveal some characteristic differences in wind-aloft patterns. One profile is characteristic of a well-mixed atmosphere without distinct layers.

In another wind shear occurs in a region of abrupt change in windspeed, and in another where there is a sharp change in direction. An interesting feature of another profile is the occurrence of a low-level jet wind near the surface with relatively low windspeeds above.

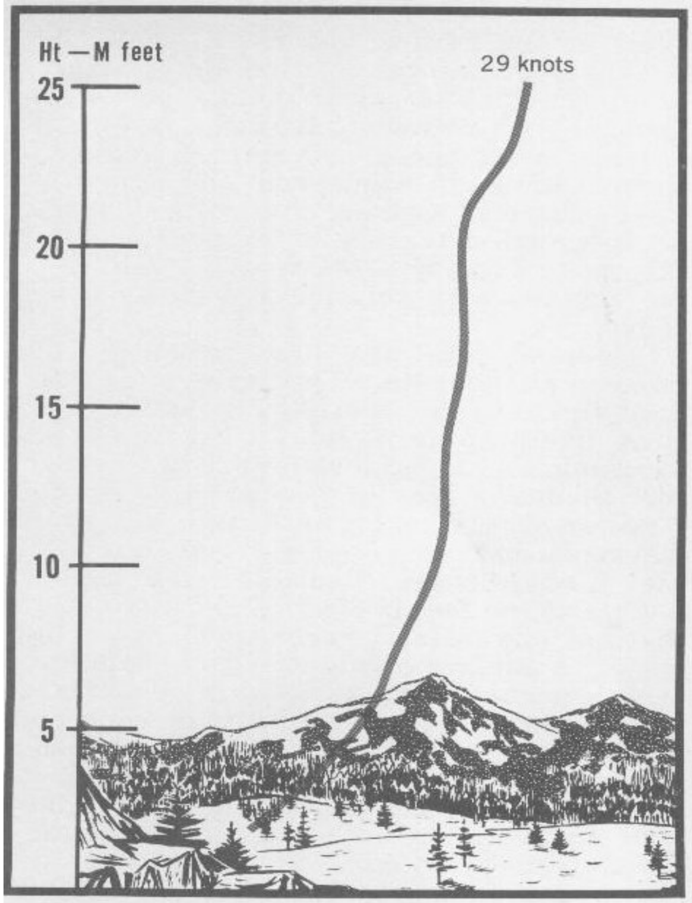
Low-level jets are predominantly mid-western phenomena. This is not to say that they do not occur in other areas. They are most likely to form where conditions are favorable to a layered structure in the lower few thousand feet of atmosphere. In fair weather this strongly suggests a marked difference between day and night probabilities of occurrence. Stratification in the first few thousand feet is discouraged by daytime thermal mixing and encouraged by cooling from the surface at night. These jets have been observed, for example, to reach maximum speeds in the region just above a nocturnal inversion. They have not been studied in rough mountain topography; however, the higher peaks and ridges above lowland nocturnal inversions may occasionally be subjected to them. The geographic extent over which any one low-level jet might occur has not been determined.

In the vicinity of the tropopause, the transition zone between the troposphere and the stratosphere, there are a number of belts of strong winds known as jetstreams. These are belts of strong westerly winds that circle the hemisphere in often meandering discontinuous segments. Each segment may be a few thousand miles in length, up to perhaps 400 miles in width, and often 4 to 5 miles in depth. Segments are sometimes broken into separate bands several hundred miles apart. The jetstream swings farther south in winter, moving north again in the summer. Speeds in the jet core have been observed up to 250 knots or more, and speeds of 140 knots or more are common.

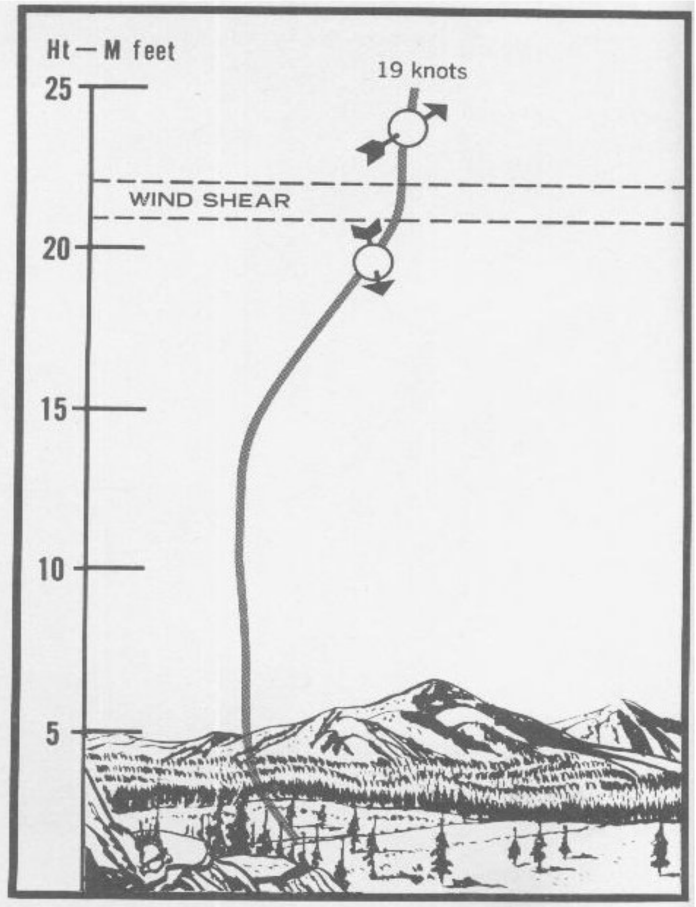
General Winds Near The Surface

General winds near the earth’s surface are caused by the general circulation winds aloft. “General winds” in this sense distinguishes surface winds related to the general circulation from the more local convective winds discussed under that heading. Although generated b y the winds aloft, the general wind is altered considerably in both speed and direction depending on roughness of the surface and the presence or absence of convective mixing. Surface wind direction is indicated on weather maps by a tailed circle representing a wind arrow. The circle represents the head and the arrow “flies” with the wind.

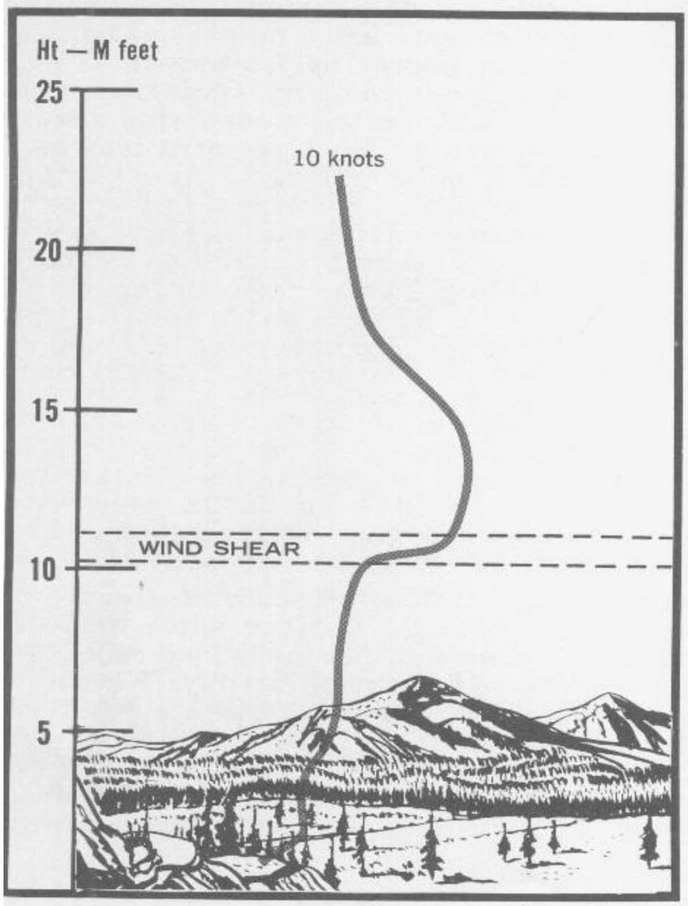
In descending from aloft there is a transition in both wind speed and direction until the surface is



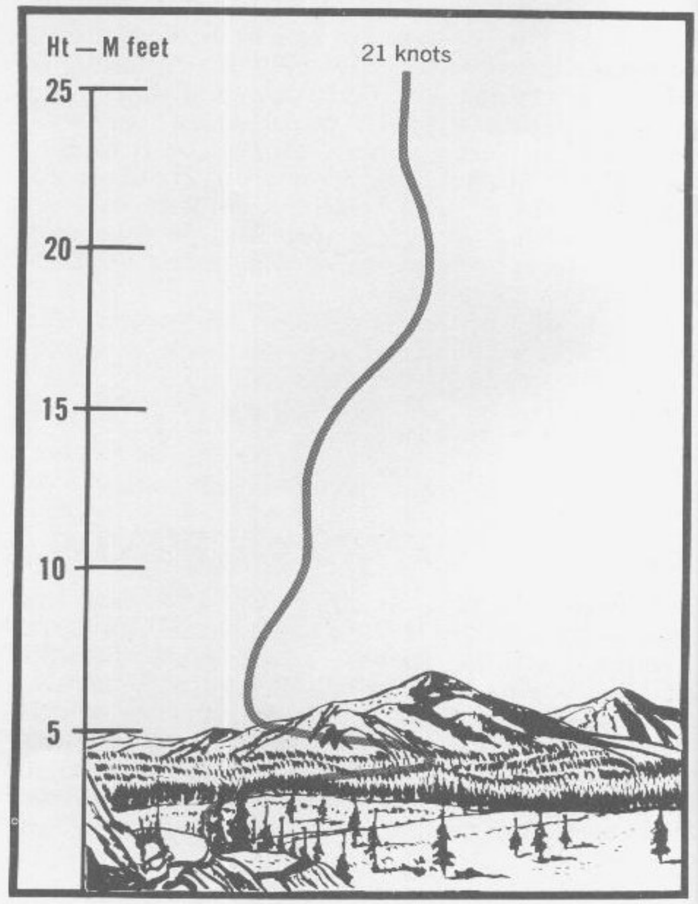
Gradually increasing windspeeds aloft.



Change in direction may cause wind shear.

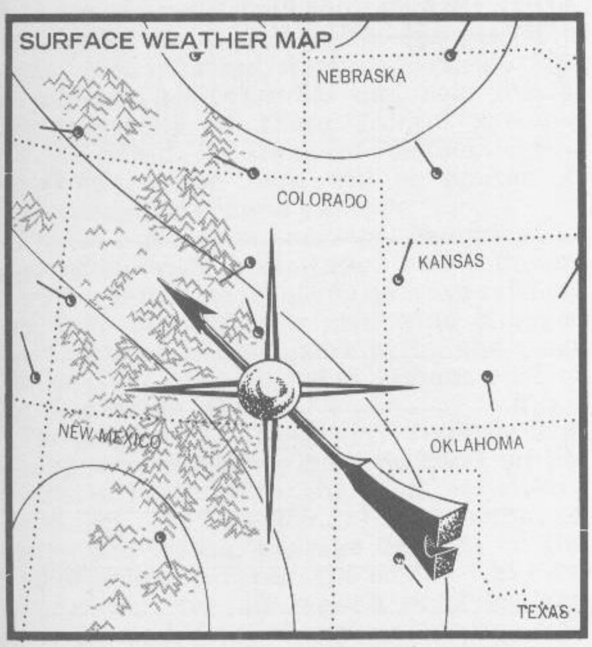


Shear occurs where windspeed changes rapidly.



Low-level jets appear most commonly at night.

reached. The nature of the transition depends on the roughness of the terrain below, presence or absence of instability in the lower atmosphere, and depth of the unstable region. All of these vary widely with time and between localities.



Wind directions at surface stations may differ widely from the wind above the mixing layer.

Surface Friction and Stability

Friction with the earth's surface, in addition to causing turbulence, acts to slow down air movement near the surface and by so doing also affects local direction. The rougher the local surface, the more pronounced these effects become. The effect of the earth's rotation in causing wind to bend to the right is offset in part by surface friction which bends it back to the left. Thus, winds near the surface tend to spiral diagonally outward across the isobars around the HIGHS and inward toward the centers of LOWS.

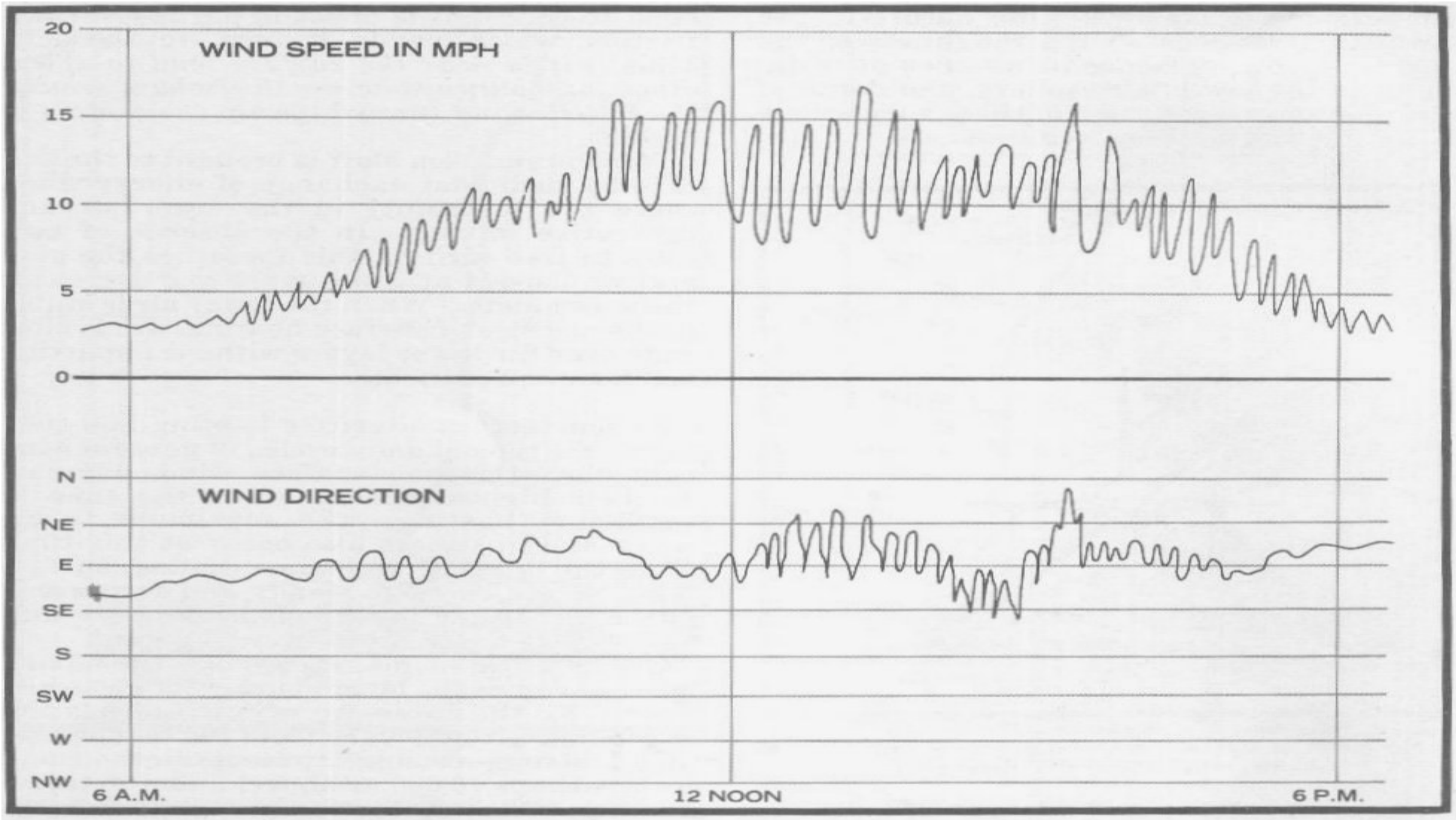
Momentum from aloft is brought to the surface by turbulent exchange of energy when there are instability in the lower air and convective mixing. In the absence of barriers to free airflow, this increases the general windspeed at the surface and decreases the speed aloft. When the lower air is stable in the absence of surface heating, winds aloft slide over the lower layers without imparting much motion to them.

In flat terrain alternate heating and cooling cause typical daily cycles in general wind behavior. Daytime surface winds increase to their highest speeds about the time of maximum heating. The maximum turbulence and gustiness also occur at this time. With the onset of nighttime cooling, surface winds again begin to steady and decrease in speed. If a low inversion forms, they may decline to a lazy drift or even to a calm.

Depth of the instability region-the mixing layer-over open, level country is governed mostly by the intensity of surface heating or cooling. It may vary from 100 feet or less, under strong cooling, to several thousand feet, perhaps 10,000 or above, under extreme heating. With stable surface air, mixing is limited in depth to the surface layer in which there is mechanical turbulence. It may disappear entirely with low windspeeds beneath a surface inversion.

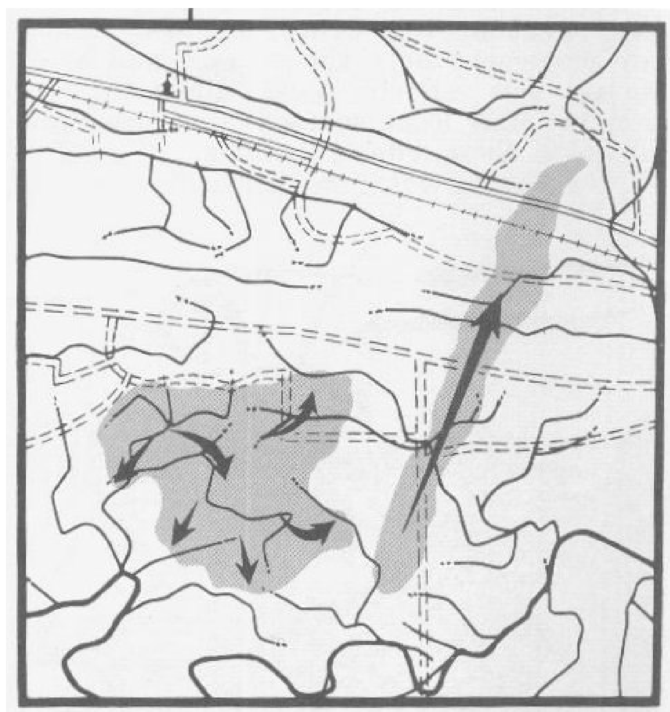
In mountainous areas the effect of the mixing layer on surface windspeed is more complex. The mountain peaks and exposed ridges may be in one regime while the valleys may be in a different one. This is particularly true when nighttime cooling forms inversions part way up the slopes.

Surface heating causes thermal turbulence and mixing through a deep layer.



Alternate heating and cooling over flat terrain causes cyclic changes in wind behavior.

Even in the absence of significant mixing, true laminar flow is probably rare in the atmosphere, and particularly so in the general winds near the surface. But times occur in stable air flowing over relatively smooth surfaces when turbulence is only minor. Then, for all practical purposes, surface winds as well as the winds aloft have the steady speed and direction characteristic of laminar motion. For example, while turbulent winds usually cause more erratic fire behavior, the laminar type often result in more rapid and sustained fire runs in one direction. Open plains and gently rolling topography most frequently experience general winds of the laminar type.


Nature of the wind during a wildfire is shown by shape of the burned area.

Frontal Activity

The frequency of change in general windflow associated with the general circulation is somewhat greater in eastern portions of the continent than in the mountainous West. The East experiences more frequent and rapid movement of pressure systems than occur during much of the year in the West.

In the West, the major mountain systems tend both to hinder the movement of high-and low-pressure areas and to lift winds associated with them above much of the topography.

As successive air masses move across the land, the change from one to another at any given point is marked by the passage of a front. Two frontal systems in particular are of interest: warm fronts and cold fronts. Characteristics of these frontal systems are described in standard meteorological texts. Discussion here will be limited to wind changes as fronts pass by.

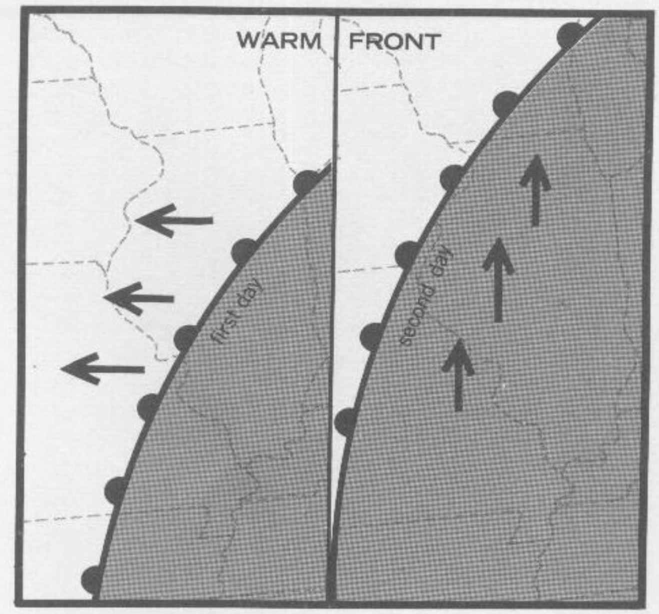
Fronts are most commonly thought of in association with precipitation and thunderstorms. But some fronts do not cause either; and they in particular give wind changes their particular significance to fire and other fair-weather phenomena in regions east of the continental Divide.

The passage of a front is invariably accompanied by a shift in wind direction. In the northern hemisphere the shift is always clockwise. The particular wind behavior during the front passage depends on the kind of front, its speed, the contrast in temperatures of the air masses involved, and on local conditions of surface heating and topography.

When a warm front passes, surface winds shift usually from 45° to 90°. This usually means from east or southeast to south or possibly southwest. Steady winds are the rule both ahead and behind the front, since the surface air is relatively stable. Surface winds are often gentle to moderate, and the shift is usually gradual.

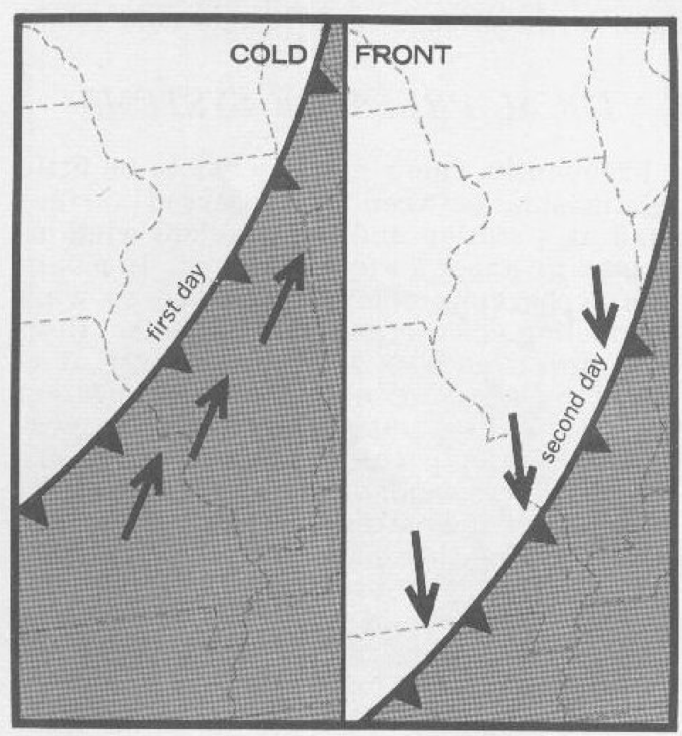
But the passage of a cold front is different. The change is usually sharp and distinct, even when the air is predominately dry and without strong temperature contrasts. Ahead of a cold front the wind ordinarily blows from some quarters on the left when facing the front. East of the continental Divide this usually means from the south or southwest. As the front approaches, wind typically increases in speed and often becomes quite turbulent. If cold air aloft overruns warm air ahead of the surface front, the resulting instability may cause this turbulence to be violent.

The wind shift with passage of a cold front is abrupt and may be from 45° to 180°, with the wind then blowing from the northwest or north after the front has passed. Gustiness may prevail for some time after the frontal passage, since the cooler air flowing over warmer ground may become unstable. If the temperature contrast is not great, however the winds soon steady and may be relatively gentle.

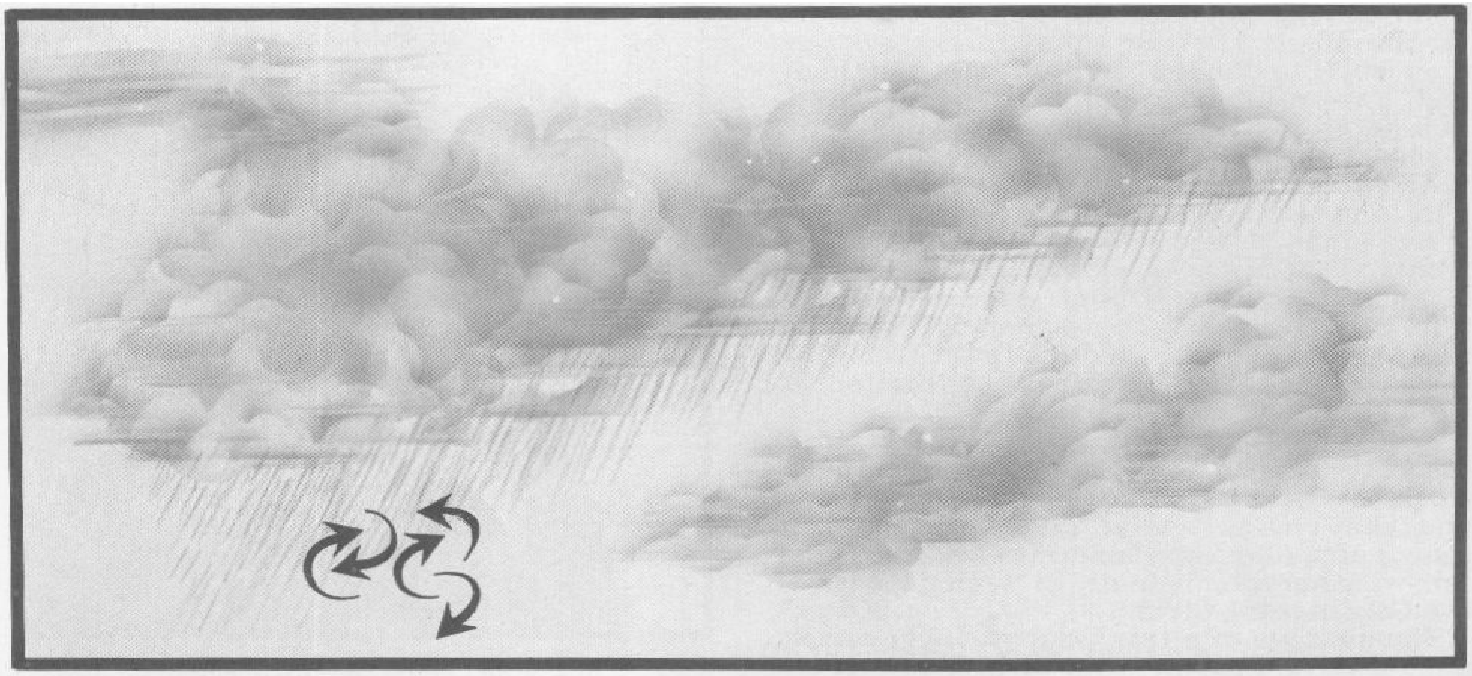


Wind is steady and shifts gradually clockwise when a warm front passes.

Cold fronts are often preceded by squall lines. These are narrow zones of instability that often form ahead of and parallel to the front, but occasionally form elsewhere. Squall lines associated with severe lightening storms in the Midwest sometimes have extremely violent surface winds. In less severe situations, they strike quickly with gusty winds, but in most cases last only minutes.



Wind increases ahead of a cold front, becomes gusty, and shifts rapidly when the front passes.



Squall lines produce violently turbulent winds for a few minutes.

Winds then revert to the speeds and directions they had prior to the squall.

Squall lines are most frequently narrow bands of thunderstorms and heavy rain. These storms are sometimes well scattered along the line, however, so that any one local area might experience squall-line wind behavior without rain or lightning.

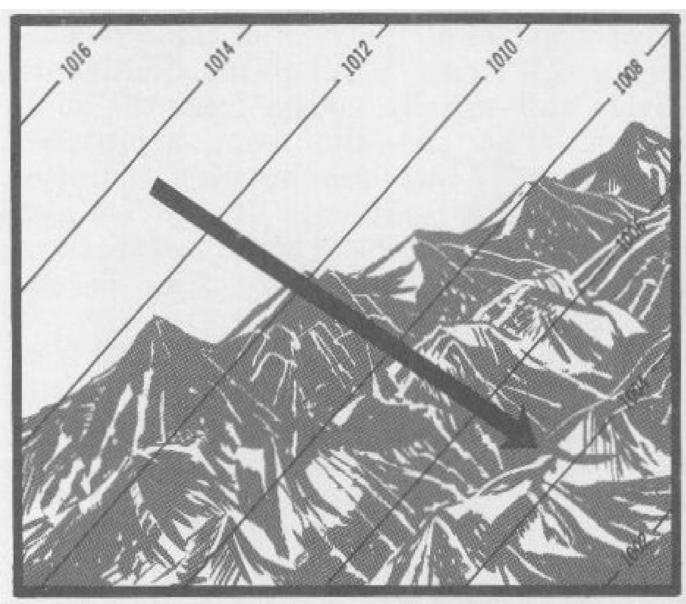
An occlusion occurs when a cold front overtakes a warm front. The wind shift accompanying an occlusion is usually 90° or more, but ordinarily more gradual and less violent than with passage of a simple cold front.

Local Pressure Systems

Frequently, there may appear to be little relationship between the observed surface wind at a station and the gradient wind indicated by upper air observations. In mountain topography this may be due to wind channeling or other mountain effects. Both here and over level terrain, however, it is often the result of a smaller configuration pressure system superimposed on the general circulation pattern. These systems are frequently too small or too brief to affect the large-scale wind pattern, or even to be identified on the weather map. Nevertheless, they may cause important locals winds.

Local, but steep, pressure gradients may result from local mechanical compression associated with the large-scale circulation. Convergence of two airstreams blowing together at an angle causes a local pressure rise in the convergence zone. These streams may have been channeled by the topography or result from waves or bulges in the general wind pattern. Mountain ranges often separate air masses having different temperature and density characteristics. Local pressure gradients caused by differences in heating of adjacent surfaces are discussed separately under "Convective Winds."

Over short distances, winds of considerable speed may blow directly across the isobars from higher to lower pressures. The tendency to bend caused by the earth's rotation and by friction is negligible. The flow is often a direct density flow and may be guided by topography into the principal



Local pressure differences cause airflow across the isobars.

drainage channels. Winds of this nature are common in both coastal and inland mountain regions. When these winds cross mountain ridges, they sometimes blow as small-scale foehn winds on the lee slopes.

More commonly, where temperature contrasts are strong the wind behavior is that of a small-scale cold front. There is much turbulence with gusty winds at the leading edge and abrupt change in the wind direction as the front passes. In all cases the local wind behavior near the surface depends on the particular distribution of pressure gradients involved, surface and upper air temperature patterns, and other local factors.

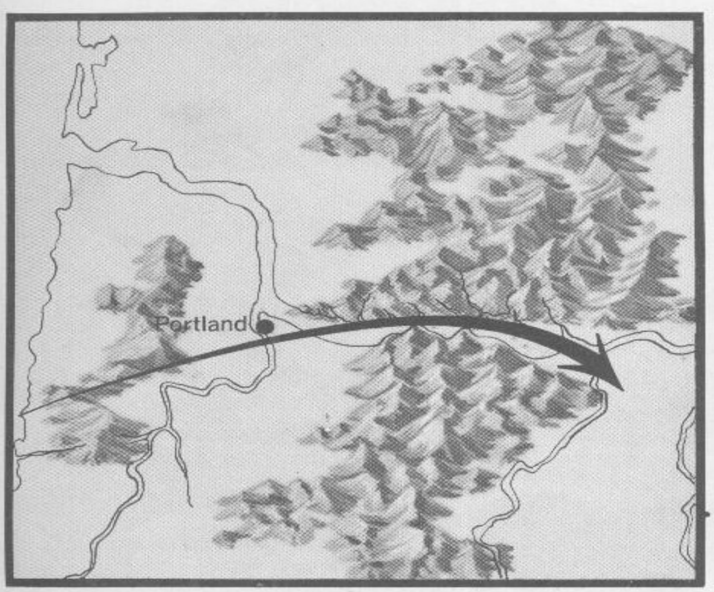
Mountain Topography

The effects of heating and frontal activity on general windflow differ somewhat between level terrain and mountain topography. Mountains represent the maximum degree of surface roughness and thus provide the greatest friction to large-scale surface circulation. Mountain chains also are effective as solid barriers against overland airflow, particularly dry cold air of polar origin and relatively cool summertime Pacific marine air. Although warm air may rise and flow over the tops, cool surface air is often either held back or deflected by major mountain systems. In addition to these mechanical effects, strong convective activity in mountain areas often damps or replaces the general wind felt at the surface. Thus, it is in the absence of strong surface heating that the general winds are often most pronounced. In this case the mountains and their associated valleys provide important channels that establish local wind behavior.

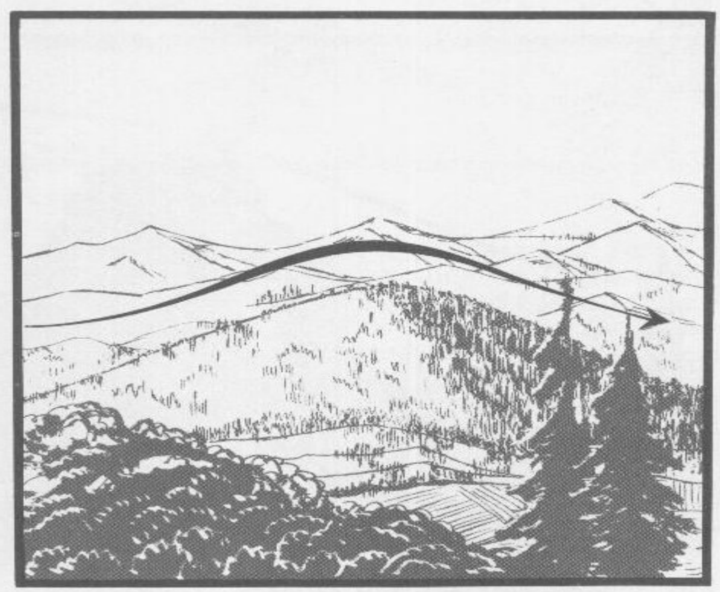
General winds blowing across mountain ridges are lifted along the surface to the crest. If the air is stable, it will increase in speed as it crosses the ridge. Ridgetop winds thus tend to be somewhat stronger than winds in the free air at the same level.

How the air behaves on crossing a ridge is influenced by ridge shape and wind speed and direction. Round-topped ridges tend to disturb surface airflow the least. In light to moderate winds there is often little evidence of any marked turbulence. Sharp ridges, on the other hand, nearly always produce a significant amount of turbulence and eddies on the lee side. Some of this is felt at the surface as gusts and eddies for short distances below the ridgetop, while much of it continues downwind aloft. Wind blowing perpendicular to the ridge line results in the least complex wind structure downwind. Most of the eddies formed are of the roll or horizontal type. If the angle of wind approach deviates from the perpendicular by some critical amount, perhaps 30° or less, vertical eddies are likely to be found in the lee draws below the ridgetop, in addition to eddies in other planes.

Eddy currents are often associated with bluffs and similarly shaped canyon rims. When a bluff faces downwind, air on the less side is protected from the direct force of the wind flowing over the rim. If the wind is persistent, however, it may start to



Mountains channel the principal low-level winds through the valleys.



Laminar flow may continue over rounded hills at low windspeeds.

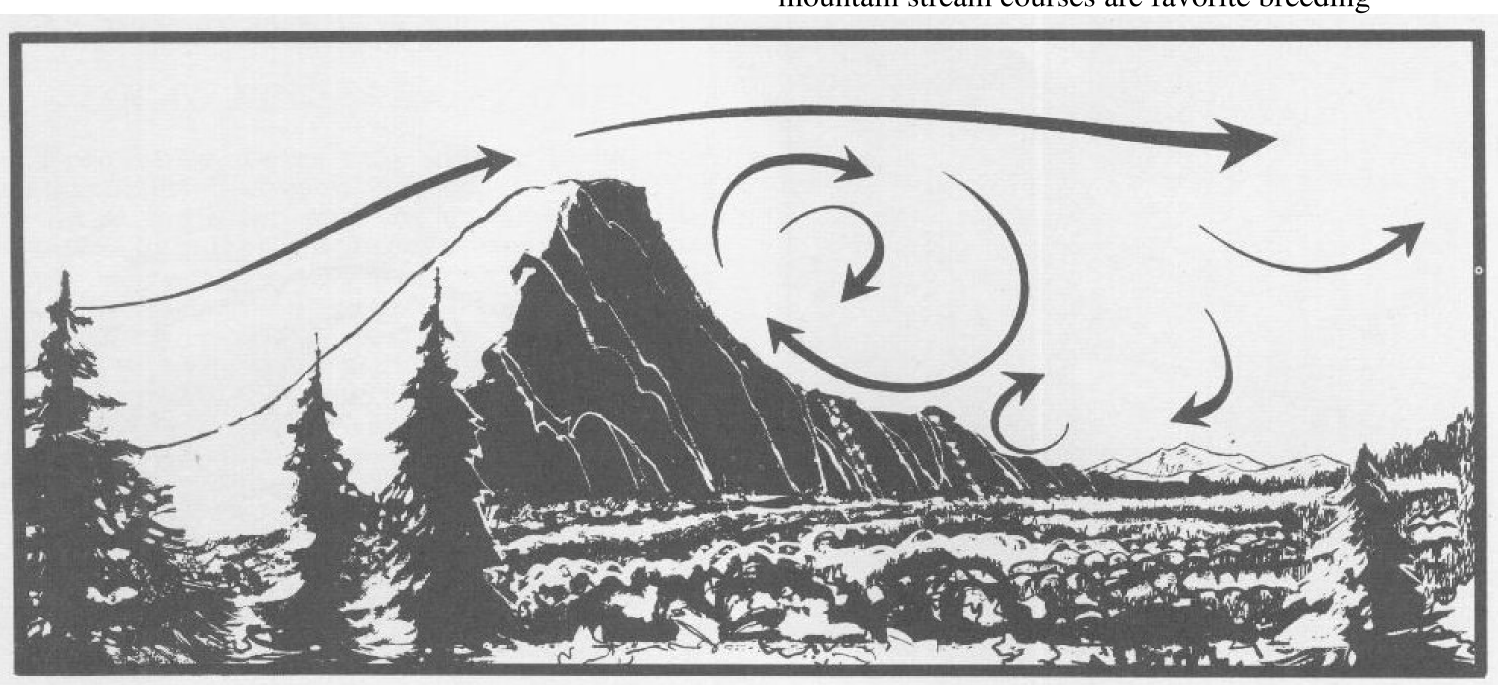
Higher windspeeds and sharp ridges cause turbulence in the lee.

rotate the air below and form a large, stationary roll eddy. This often results in a moderate to strong upslope wind opposite in direction to that flowing over the rim. Eddies of this nature are common in the lee of ridges that break off abruptly, and beneath the rims of plateaus and canyon walls.

Ridgetop saddles and mountain passes form important channels for local pressure gradient winds. Flow converges here as it does across ridgetops, with accompanying increase in windspeed. After passing through mountain saddles the wind often exhibits two types of eddy motion on the lee side. One takes the form of horizontal eddies rolling or tumbling down the lee slope or canyon, although the main eddy may be stationary. The other is more usually a stationary vertical eddy in one of the sheltered areas on either side of the saddle. Some of the these vertical eddies may also move on downwind.

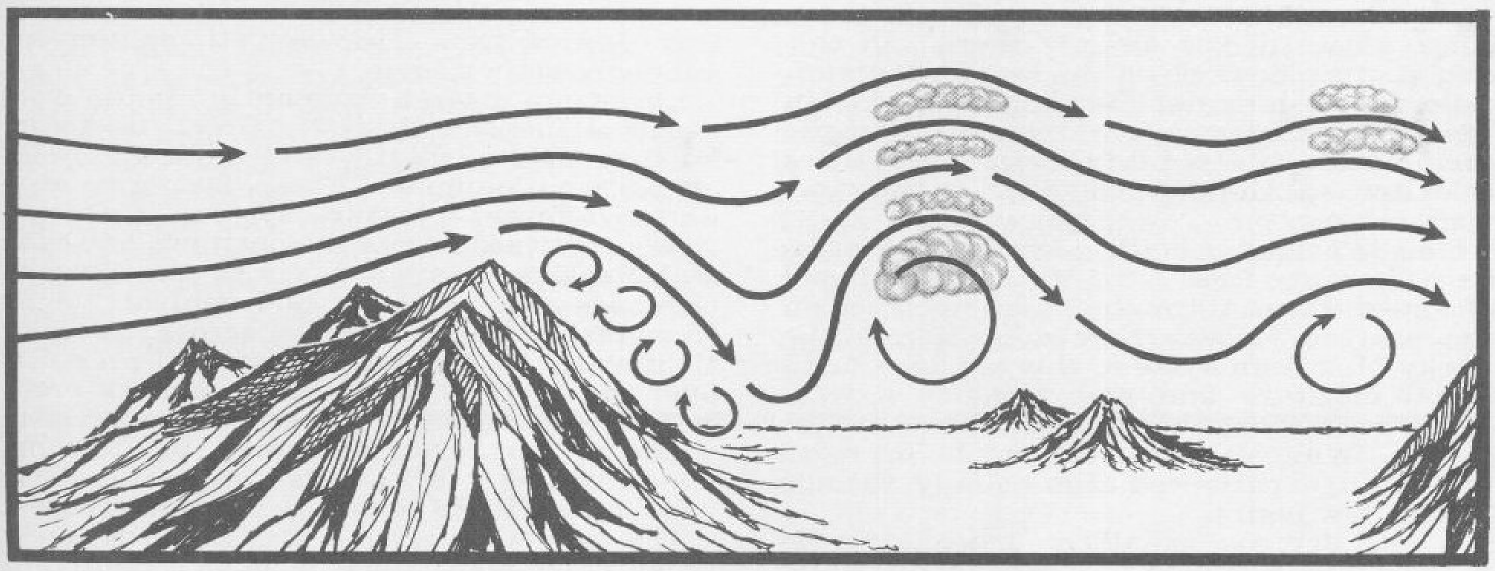
Moderate to strong winds blowing across high mountain ranges may cause large-scale turbulence for several miles downwind from the crest. In addition to the small eddies rolling down the lee slopes, one very large roll eddy may form over the lee valley with succeeding smaller ones still farther downwind. Over each of these is a lee wave or standing wave with strong updrafts and downdrafts extending thousands of feet in depth. These may be as high as 40,000 feet or above in the best known Bishop Wave in California. Large waves occur in the Rocky Mountains and on a lesser scale in the Appalachians and elsewhere. The large roll eddies may be topped by roll clouds and the individual waves by cap clouds.

General winds that are channeled in mountain canyons are characteristically turbulent. The moving air in canyon winds is in contact with the maximum area of land surface. Alternating tributaries and lateral ridges contribute toward maximum roughness. Whether the canyon bottom is straight or crooked also has an important influence on the amount of turbulence to be expected. Sharp bends in mountain stream courses are favorite breeding



Large roll eddies are typical of bluffs or canyon rims.

Mountain saddles generate typical eddies.



Lee waves have strong updrafts and downdrafts.

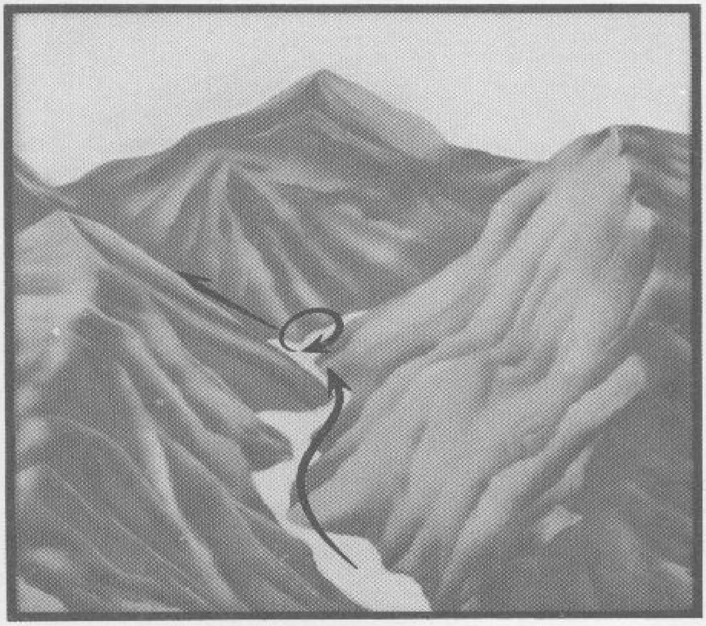
Grounds for large eddies, particularly where the canyon widens to admit a side tributary. Such eddies may be a half mile or more in diameter. They are most pronounced near the canyon floor and dissipate well below the ridgetops.

Foehn Winds

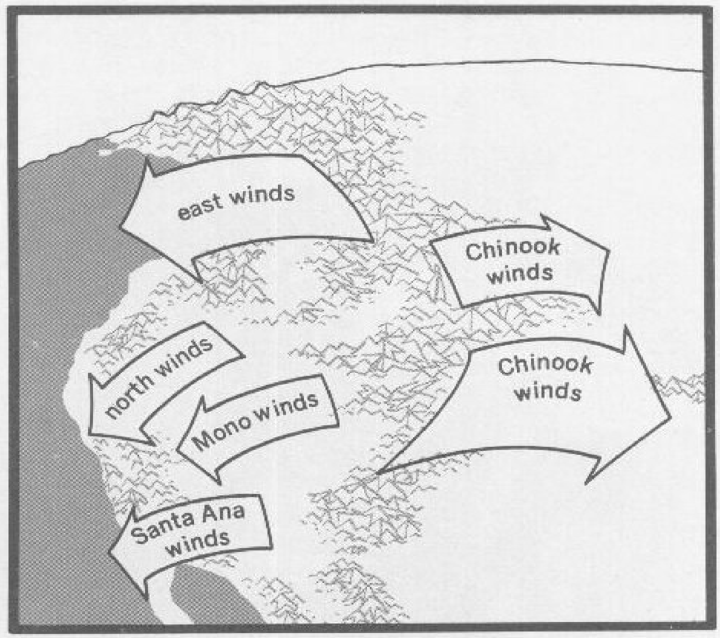
Circulation patterns in parts of the West are sometimes interrupted by the occurrence of a foehn. A foehn is a dry downslope wind, characteristic of most mountain areas. Its full development requires a strong high pressure system and a corresponding well-situated LOW.

Mostly restricted to the cool months, September to April, two types of foehn winds are common in western America. One results from air losing its moisture when forced across a major mountain chain. The other results from the flow of initially cold dry air from a higher to lower elevation. The distinctive properties common to both of these winds are that they blow downhill, are warm, and become progressively more desiccating as they descend.

Foehn winds of the first origin result when a deep system of moist air forced upward and across a mountain range is cooled and loses much of its moisture. On the lee side of the range it may then be cooler, drier, and more dense than the air already



Large persistent eddies form in bends in mountain canyons.



Foehn winds are common in these regions.

there. In this case it often flows down the lee slope, under-riding or pushing out the old air ahead of it. This downflowing air warms by adiabatic compression and is often warmer and drier than it was at the same elevation on the windward slopes.

Moist pacific air flowing across the Sierra-Cascade range loses some of its moisture and often exhibits mild foehn characteristics on the eastern slopes. Forced on across the Rocky Mountain system, this air loses additional moisture and may produce a well-developed foehn on the eastern slopes in that region. Windspeeds of 20 m.p.h. or more are often experienced, but usually subside after a few hours.

A cold, dry, and usually stagnated air mass is the second common source of foehn winds. Mountain barriers coupled with a particular distribution of high- and low-pressure areas often cause cold air masses to pile up and stagnate. Such a mound of cold air can persist until deep enough to spill over the mountain barrier, or until unblocked by some change in the large0scale pressure pattern. When released, this cold dry air flows downward as a foehn because of its density. To produce these foehn winds, massive topography undoubtedly plays an important role in helping to stagnate and pile up cold air masses.

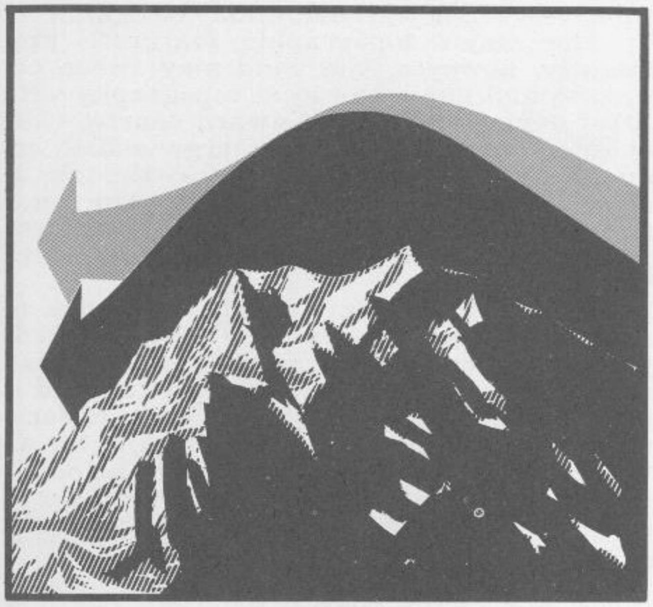
Surface windspeeds from 40 to 60 m.p.h. are common in a foehn of stagnated air-mass origin, and speeds up to 90 m.p.h. have been reported. The wind often lasts for 3 days or longer with gradual weakening after the first day or two. They sometimes stop as suddenly as they begin.

Large air masses frequently stagnate in the cool months over the Great Basin in Western United States and adjacent areas of British Columbia. These may give rise under different pressure systems to foehn winds eastward across the northern and central Rockies, westward across the Oregon Cascades and northern and central Sierra Nevadas, or southwestward across southern California. A combination of high pressure over the State of Washington and low pressure in the Sacramento Valley causes north winds in northern California. Brief foehn winds lasting 1 or 2 days may result from migratory HIGHS following the same routes.

The course of a foehn may be either on a broad front many miles wide or a relatively narrow, sharply defined belt, depending on its source and on the local atmospheric situation.

A foehn, even though it may be warm, often replaces the air ahead and occupies the surface on the adjoining plains or lowlands. Counter forces, however, sometimes prevent this and cause the foehn to override the existing system and thus not be felt at the surface at the lower elevations. At other times, a foehn may reach the surface only intermittently at scattered points, causing short-period fluctuations in local weather. Any of several mechanisms can cause these variations in foehn behavior.

The Chinook, a foehn wind on the eastern slopes of the Rocky Mountains, may replace cold continental air in Alberta and on the Great Plains.



A strong foehn may follow the surface; weaker may override the air below.

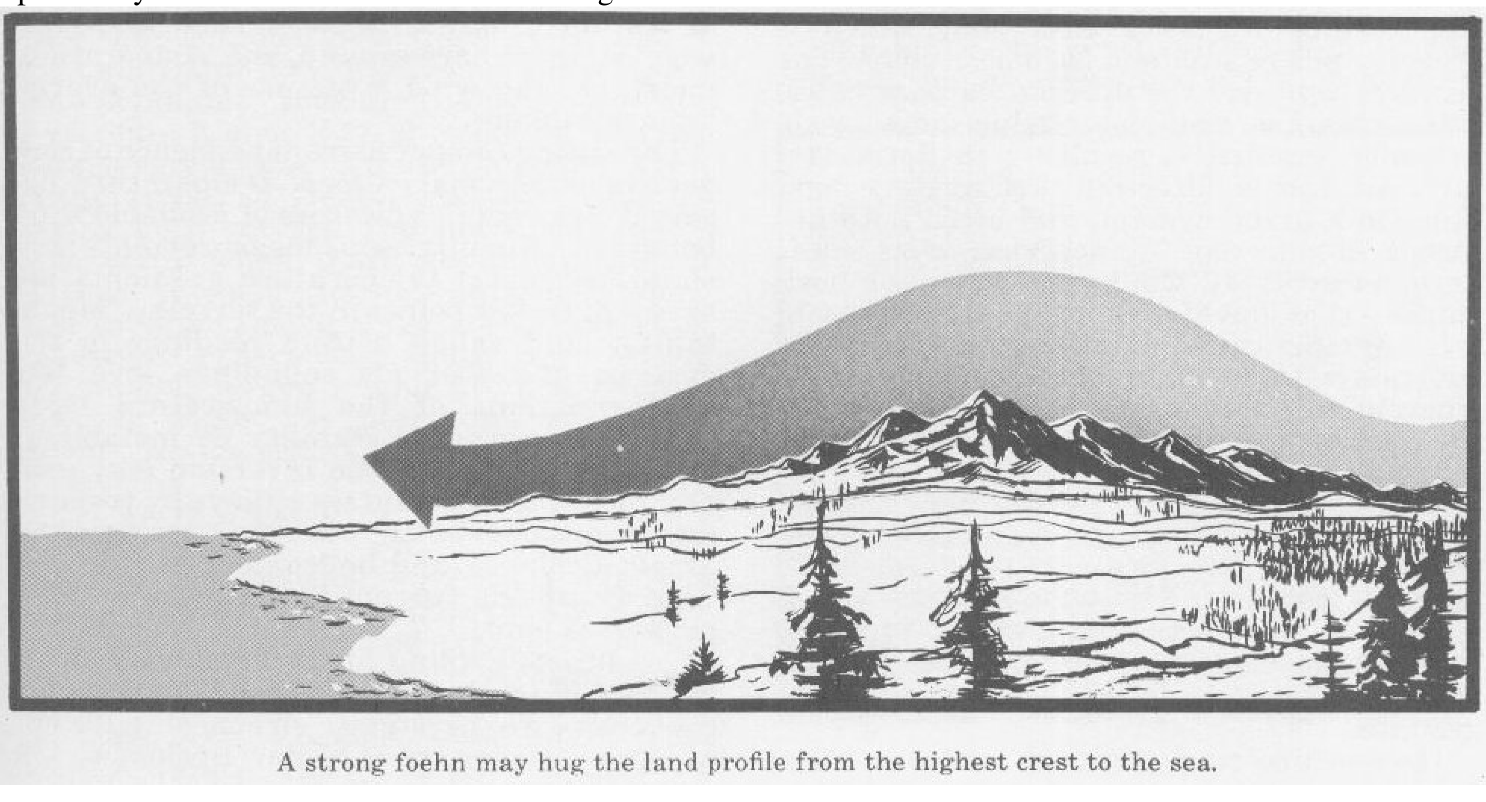
Quick wintertime thawing and rapid snow evaporation are characteristic. If the cold air is held in place by the local pressure and circulation system, the foehn will override it; or if the cold air tends to stay on the bottoms because of its greater density, the Chinook may reach the surface only in the higher spots. Chinook relative humidities of 5 percent or less and temperature changes of 30°F to 40°F. within a few minutes are not uncommon.

Along the Pacific coast a weak foreign may be kept aloft by cool onshore maritime air. A strong well-developed foehn, by contrast may cut through all local influences and affect all slope and valley surfaces from the highest crest to the sea. East winds in the Northwest, for example, sometimes flow only part way down the lee slopes of the Cascades, then level off above the lowlands and strike only the higher peaks and ridges of the coastal mountains. At other times virtually all areas are affected.

North and Mono winds often develop in northern California under conditions favorable to their flow along the surface, at least as far as the western edge of the central valley. While the path of any one event may be confined to a relatively narrow channel, it is ore or less on the surface throughout its length. Both of these winds are most common in late fall.

Santa Anas in southern California vary widely in individual behavior. One might flow seaward as a broad fan across the coastal region; another as a narrow ribbon slicing across the topography; or another in multiple, separated tracks. Their paths may touch all surfaces en route or affect only the higher elevations. All are usually characterized by their mild temperatures, extremely low relative humidities both day and night, high speeds, and strong gusts and eddies. These winds sometimes generate standing waves as they cross the southern California mountains.

The trajectory of a Santa Ana is sometimes made visible over the inland desert regions by the



A strong foehn may hug the land profile from the highest crest to the sea.

dust picked up en route. On the coastal side of the mountains the course can sometimes be seen as a band of relatively clear air cutting through a region of otherwise limited visibility.

The local behavior of a Santa Ana depends on whether it is "strong" or "weak." The relative strength is determined by size and depth of the air mass buildup in the Great Basin, surface pressure gradient to the sea, and temperature and density contrasts between the Santa Ana air and the air it penetrates or displaces.

Typically in southern California during the Santa Ana season, there is a daytime onshore breeze with gentle to weak up-slope and upcanyon winds in the adjacent mountain areas. With nighttime cooling these directions reverse to produce down-canyon and offshore winds, usually of lesser magnitude than the daytime breeze.

A strong Santa Ana wind wipes out this pattern. It strikes the upper desert-facing mountain slopes, blows up and through the mountain slopes, blows up and through the gaps and over the ridges, then down the surfaces of leeward slopes, canyons, and more broad valleys to the sea. Occasionally, if relatively shallow, the flow may be channeled by the major topographic features. Frequently, however, the wind may sweep up, across, and down the local topography with chanical turbulence with large eddies induced by topographic features, though, is often sever. A strong Santa Ana thus sweeps out the air ahead of it and often show little or no difference in day and night behavior.

As the Santa Ana weakens, it begins to show a diurnal behavior. Its lower speeds permit appreciable warming during the day and cooling at night. This tends to hold it aloft during the day and even permits some upslope movement of the Santa Ana air on the lee slopes. At night, cooling of the Santa Ana flow permits it to reinforce the normal land-to-sea breeze, resulting in somewhat higher-than-usual downslope winds. Cooling at night is sometimes strong enough to create inversions in the coastal valleys. The flow is then held above the valley floors in the late night hours. As the Santa Ana dies, marine air flows back and the normal daily cycle is resumed.

Convective Winds

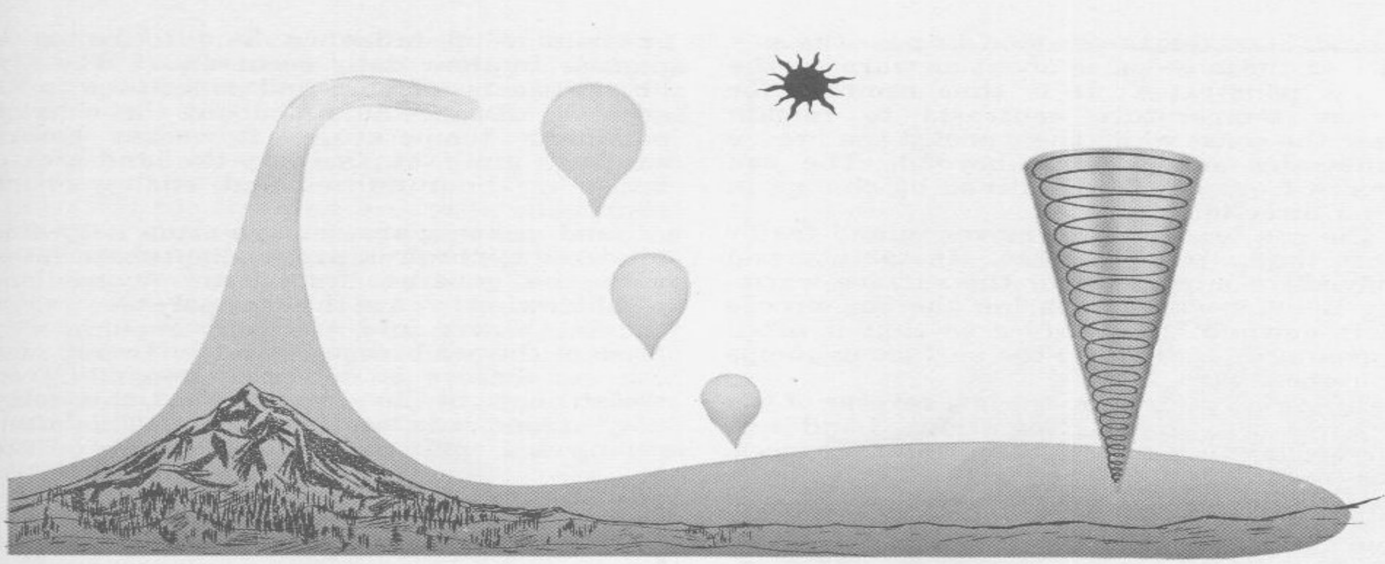
In the absence of strong pressure gradients aloft, local circulation in the atmosphere is often dominated by currents resulting from temperature differences within the locality. Air made unstable by warming at the surface tends to rise; that which is cooled tends to sink and become stable. The surface temperature differences that cause this vertical motion also produce small-scale pressure gradients, resulting in horizontal airflow. Either direction of flow may dominate in a given system, and often both are inseparably mixed. Hence, convective winds here refers to all winds—up, down, or horizontal—that have their principal origins in local temperature differences. This is in contrast with common meteorological usage, wherein convection implies circulation with the vertical component in the upward direction only.

Various convective wind features may be augmented, opposed, or eliminated by the winds aloft or their associated general winds near the surface. The term gradient winds is used in this section to identify combinations of these surface and upper winds having their origins in the larger pressure systems.

The nature and strength of convective winds vary with many factors. Since they are temperature dependent, all features of the environment that affect heating and cooling are significant. Among the more important are season, diurnal changes, nature of the terrain and its cover such as water, vegetation, or bare ground, and temperature, moisture, and wind structure of the overlying atmosphere.

The strong temperature dependence of convective winds makes local temperature observations useful indicators of probably wind behavior. Simultaneous measurements may show horizontal temperature gradients between different points in the terrain. Mountaintop and valley bottom readings in the absence of upper air soundings give fair approximations of the temperature lapse rate and associated stability or instability. Height of the nighttime inversion may usually be found in mountain valleys by traversing side slopes with portable instruments. Small airplanes and helicopters can also be used to provide current information on temperatures aloft.

Conditions leading to strong surface heating result in the most varied and complex convective wind systems. Warm air adjacent to heated slopes tend to flow upslope to the crest where it bleeds off



Various convection systems dominate when gradient winds are weak. *Left*, continuous bleed from crest; *right*, whirlwind; *center*, intermittent bubbles.

in a more or less continuous stream. In generally flat terrain, air heated at the surface tends to build up in stagnated mounds until it reaches a critical point or is released by mechanical triggering. The escaping air may take the form of the familiar upward-spiraling whirlwind or dust devil, or in other cases, of intermittent bubbles that break off and boil aloft where they dissipate in turbulent mixing. in mountain terrain, cumulus clouds often form over the ridges, fed by the continuous supply of warm air escaping at the crests.

Air that is cooled near the surface almost invariably flows downward along the steepest route available, seeking the lowest levels. If en route it should meet colder air already there, it spreads out on top of the colder layer.

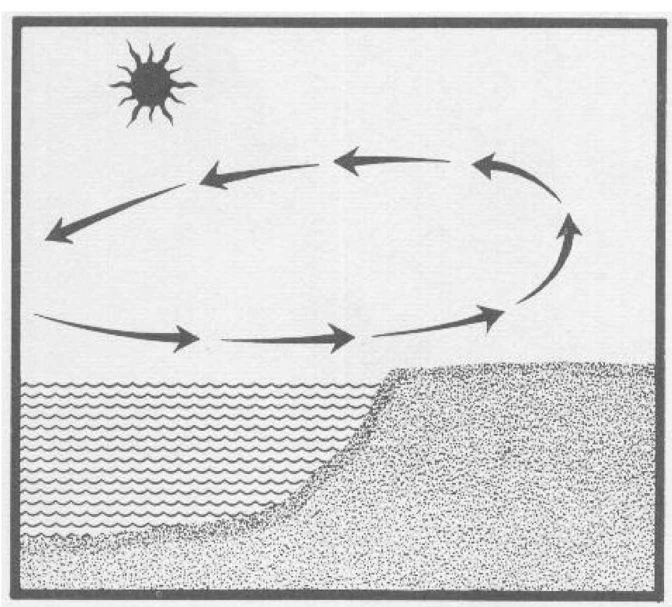
Other types of local convective winds involving both vertical and horizontal movement occur when there are differences in heating between sizable adjacent areas. Most familiar among these are the land and sea breezes experienced along ocean shores and around larger inland lakes and bays. In these winds, the horizontal wind component is usually the more significant.

Land and Sea Breezes

During the daytime when land surfaces become warmer than adjacent water surfaces, the air over the land expands, becomes less dense, and the pressure becomes lower than over the nearby water. As a result air—a sea breeze—begins to flow inland from over the water. The warm air over the land rises and cools, and on reaching higher levels tends to flow outward. With favorable gradient winds aloft, this may be from the land toward the sea where it sinks to complete the circulation cell.

The surface sea breeze starts in the morning and strengthens during the day. The breeze is felt first at the coast, but gradually pushes farther and farther inland. It may extend 30 to 40 miles or more from the water under favorable conditions.

The sea breeze brings in relatively cool, moist marine air. It is often accompanied by fog along the coast during at least the morning hours. Within the first few miles inland, however, the marine air near



The sea breeze.

the surface at times becomes about as warm as the air it penetrates. It is thus common for strong temperature contrasts to remain near the coast while the warmed sea breeze penetrates many miles beyond. The sea breeze front can be identified by change in wind direction.

The sea breeze often moves inland faster aloft than at the surface. Instability and convective mixing under the surface warming influence tend to bring the sea breeze aloft down to the surface so that it often appears to progress on the surface as jumps or surges.

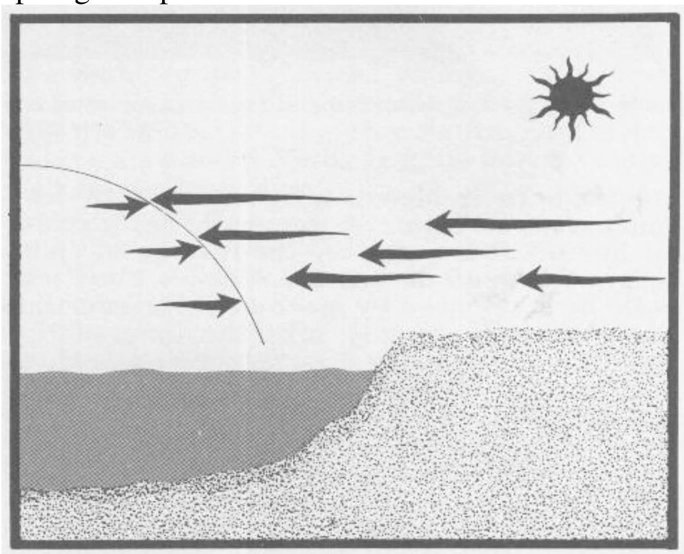
The land breeze at night is a reverse of the daytime sea breeze circulation. Land surfaces cool more quickly than water surfaces. Air in contact with the land becomes cooler at night than air over water, gains in density, and flows from the land to the water.

Flowing seaward over the cooling land, the land breeze is stable. It is consequently much more shallow than the corresponding sea breeze. The land breeze is also a more gentle air flow, usually on the order of 3 to 5 miles an hour. Return circulation to the land, if any, is likely to be lost in the prevailing winds aloft. By cooling the warmer moist marine air, the land breeze is also frequently associated with fog along the coastline.

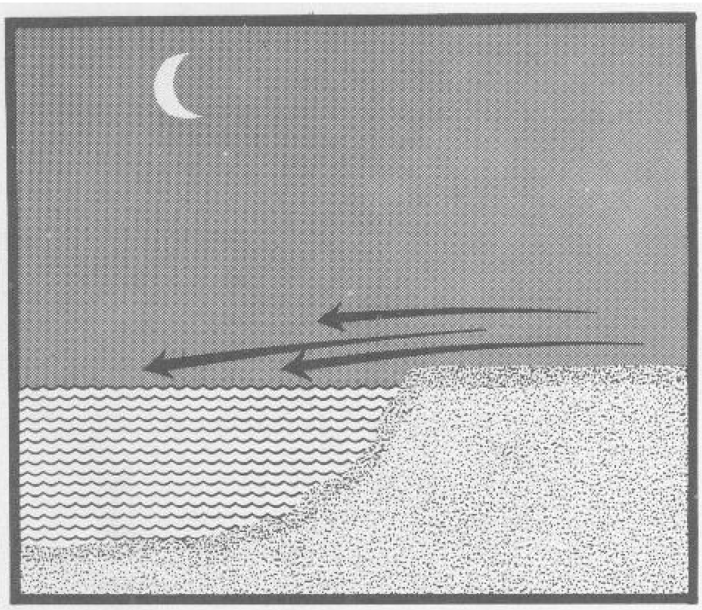
Land and sea breezes are most pronounced during the summer months, tapering off at both ends of the warm season. Whether or mot they are significant locally, though, depends on local climatic factors and on the shape and orientation of the shoreline and inland topography. They are an important feature of the summer weather along much of the Pacific coats.

Land and sea breezes in the absence of gradient wind influence tend to be quite regular in their daily occurrence. The sea breeze starts offshore and penetrates to its greatest distance inland about the time of maximum temperature. It ceases before sundown and is replaced by the land breeze beginning near sunset and ending about sunrise.

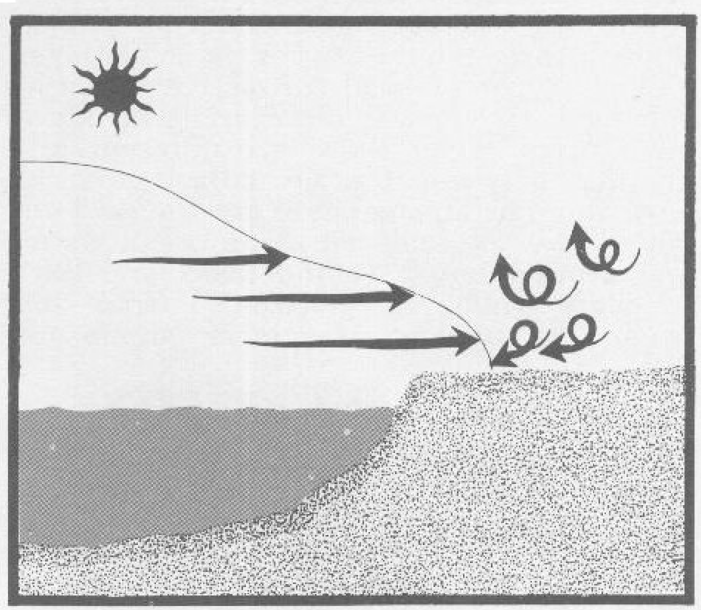
Land and sea breezes are often helped or hindered by the pressure gradients associated with the general circulation. A gradient wind blowing toward the sea operates against the sea breeze and, if strong enough, may prevent the sea breeze entirely. In any case the sea breeze is delayed. Depending on the strength of the gradient wind, this delay may extend into late afternoon. This often produces a "piling up" of marine air off the



Adverse gradient wind may block the sea breeze.



The land breeze.



A delayed sea breeze may rush inland like a cold front.

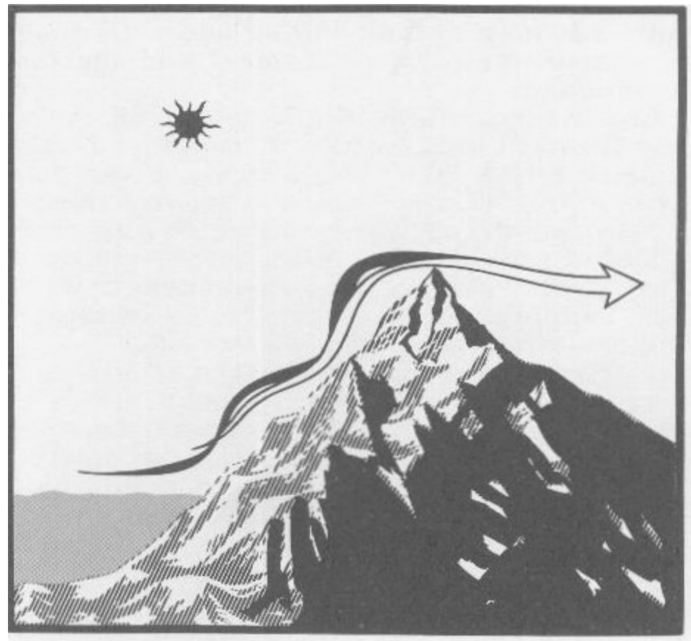
coast. If, by then, the local pressure difference has become great enough, this sea air rushes inland with characteristics of a small-scale cold front. Air behind the front is cool and moist.

The land breeze does not form against a strong onshore gradient wind. It is common, however, for the land breeze to slide under onshore winds of moderate speeds. In doing so, they do not extend very far seaward.

Gradient winds also tend to make out any closed-cell land and sea breeze circulation. With an onshore gradient wind aloft, for example, there is no return flow aloft of the daytime sea breeze. The surface air that moves inland rises, mixes with the gradient wind, and is replaced on the seaward side by gradually settling air from the general circulation

Gradient winds along an irregular or crooked coastline may amplify a land or sea breeze in one sector and oppose it in the next. Often times, too, shifting gradient winds may cause periodic reversals of these opposite effects in nearby localities, resulting in highly variable local wind patterns.

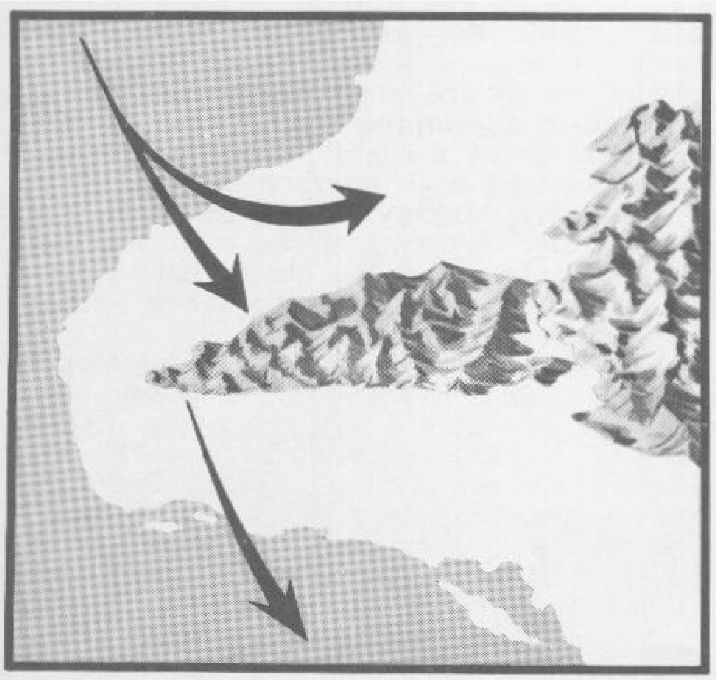
Mountains along the coastline, as along the Pacific coast, act as barriers against the free surface flow of air between the water and the land. On seaward-facing slopes the sea breeze may combine with upslope winds during the daytime. At the mountain crests, however, both mix for the most part with the air aloft and flow with the gradient winds at this level. The marine air that does flow inland over the mountains mostly stays aloft beyond the lee



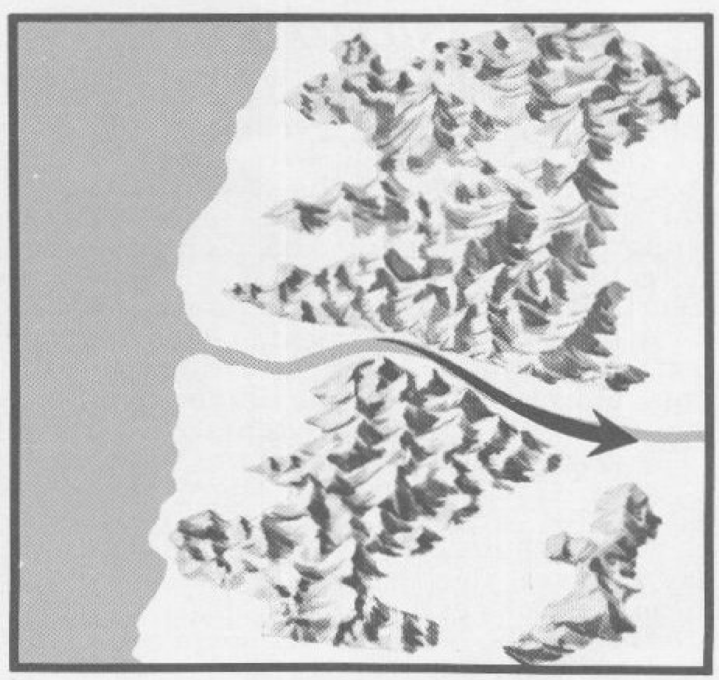
Coastal mountains restrict the sea breeze.

slopes under daytime heating without significantly affecting lee slopes and inland valleys.

River systems that penetrate the coast ranges provide the principal inland sea breeze flow routes. the flow is these is sufficient to carry tremendous amounts of marine air inland, helping to maintain inland summer humidity at a moderate level. Here the sea breeze often joins with afternoon upvalley and upcanyon winds, resulting in cool and relatively strong flow. In broad valleys this flow takes on the usual sea breeze



Gradient wind may favor a sea breeze on one sector and oppose on another.



River systems provide sea breeze channels.

characteristics, but in narrow canyons or gorges it may be both strong and very gusty as a result of both mechanical and thermal turbulence.

Coastal mountains similarly cut off major flow from the land to the sea at night. Downslope winds on the ocean-facing slopes join with a feeble land breeze from the coastal strip at night, but, again, the outflowing river systems provide the principal flow routes. The downcanyon and downvalley flow is, like the normal land breeze, a relatively shallow and low-speed wind system.

During the summer months, there is a persistent high pressure system called the North Pacific high in the general area between Hawaii and Alaska. Circulation from this is in the form of onshore winds aloft along most of the Pacific coast. These winds, although of generally modest speed, tend to reinforce the daytime sea breeze and reduce the speed of the land breeze at night. The Bermuda High in the western Atlantic Ocean has the same effect in regions bordering the Gulf of Mexico, but its offshore circulation along much of the Atlantic coast is in opposition to the daytime sea breeze and tends to limit it to a narrow coastal belt.

Small-scale diurnal circulations similar in principle to land and sea breezes occur on inland waters. lake breezes can appear along the shores of lakes or other bodies of water large enough to establish a sufficient air temperature gradient. The lake breeze is common in summer, for example, along the shores of the Great Lakes. Here, it is common on a summer afternoon for most shore stations to experience onshore winds.

Valley And Slope Winds

Winds in mountain topography are extremely complex. The general winds near the surface associated with pressure systems dominate much of the time. But whenever these winds weaken in the presence strong daytime heating and nighttime cooling, convective winds of local origin become important features of the mountain weather. These conditions are typical of clear summer weather in which there is a large diurnal range between daytime maximum and nighttime minimum surface air temperatures.

The most difficult wind systems to evaluate in mountain terrain are those in which neither the gradient wind nor convective activity are quiescent. Wind of either origin may displace, reinforce, or oppose the other. These relations can change quickly in time—often with the element of surprise. They may also differ between terrain features separated only by yards. The convective activity may dominate the surface wind structure in one instance or, through the mixing process, permit the speed and direction of winds aloft to dominate the surface flow in another apparently similar instance.

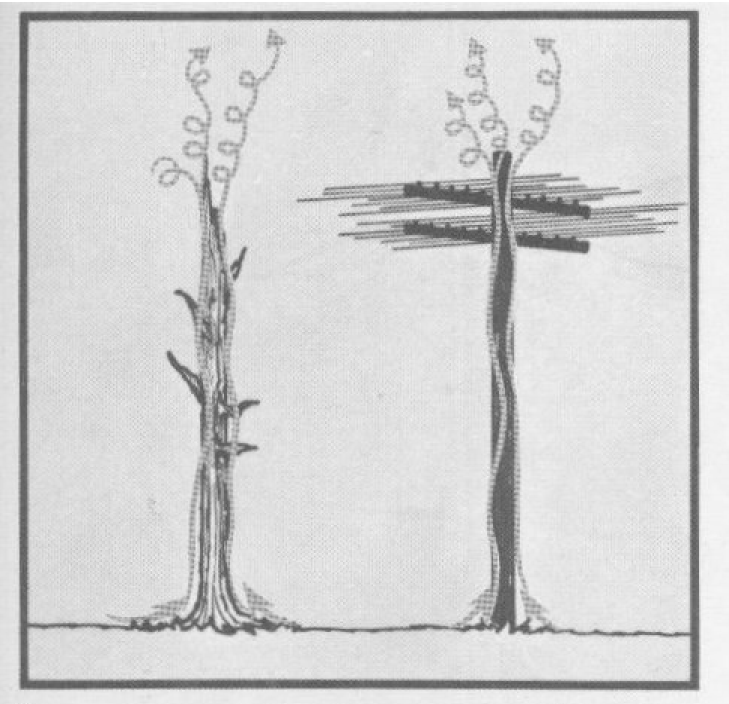
The interactions between airflow of different origins, very local pressure gradients caused by non-uniform heating of mountain shapes of mountain systems combine to prevent the rigid application of many rules of thumb about convective winds in mountain areas. Every local situation must be interpreted in terms of its uniqueness in time and space. Wind behavior described on the pages that follow is therefore typical, but subject to interruption or change at virtually any time or place.

Differences in air heating over mountain slopes, canyon bottoms, valleys and adjacent plains result in several wind systems. Although theoretically distinct in origin, these winds combine in most instances and operate together. Their common denominator is upvalley, upcanyon, upslope flow in t the daytime and downflow at night. They all result from horizontal pressure differences, changes in local air density conducive to vertical motion, or from combination of the tow.

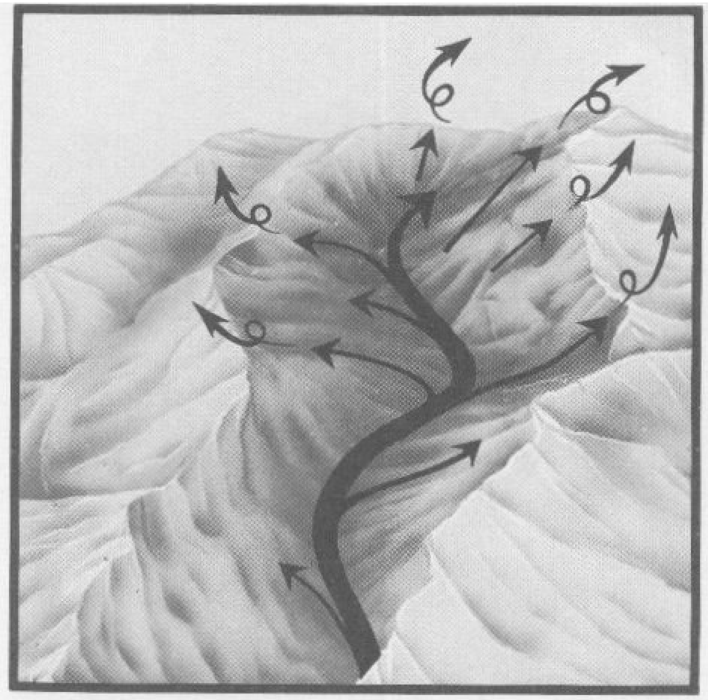
Because of the larger heating surface to which the air above is exposed, the air in mountain valleys and canyons tends to become warmer during the day than that at the same elevation over adjacent plains into which the valleys open. The larger cooling surface causes reversal of the temperature difference at night. Resulting pressure differences cause air from the plain to flow into or upvalley by day and downvalley at night. Valley winds are pronounced because they are channeled.

Slope winds are local diurnal winds that move along all sloping land surfaces. They blow upslope as a result of surface heating and downslope with surface cooling. Much of slope wind behavior can be understood most readily in terms of the effects of local heating and cooling on air density an d on the ways warm and cool air move with respect to each other.

Air that is warmed by surface heating becomes unstable. But for the warm air to rise, it must expend the energy to displace and mix the cooler air above. Thermal energy in the warm air is available for mixing once the process is started, but may have to build up to a critical value to initiate the motion. As warm air accumulates over level ground, upward



Warm air next to heated surfaces provides natural chimneys.



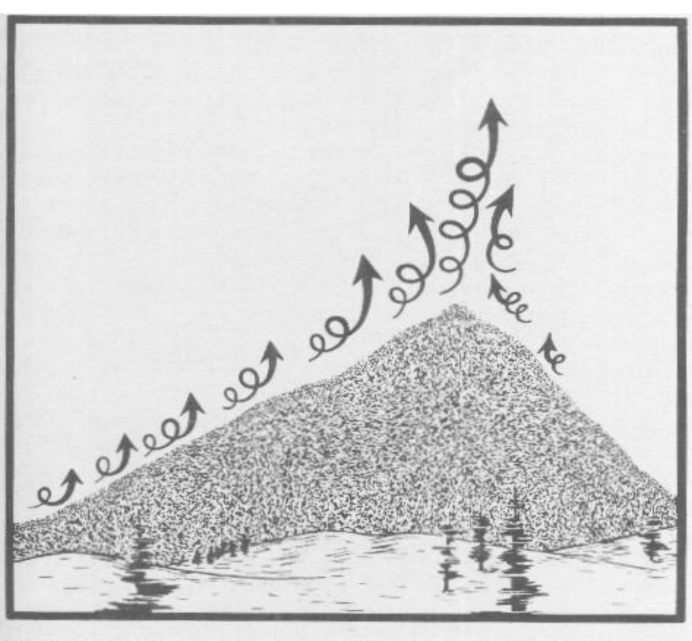
Combined valley and upslope winds exit at the ridgetops.

surges may occur at any point more or less by chance when this critical value is reached. Over sloping or vertical surfaces, however, the weak point is commonly at the highest elevation; and this is where the principal exodus takes place. This warm air sheath serves as a natural chimney, providing a path of least resistance through which additional warm air may flow.

Upslope winds in mountain terrain flow within the warmed air layer near the surface. Ravines or draws facing the sun are particularly effective chimneys because of the increased area of heating surface. Winds are frequently stronger in these than on the exposed slopes. Momentum of the upflowing air, convergence of upslope winds from opposite slopes, and mechanical turbulence in the wind at the crest may combine to facilitate the escape of hot air aloft at these upper levels. Wisps of warm air escape upward en route as turbulence in the warm air disturbs the warm-cool air boundary. This increases toward the main exit over the crest.

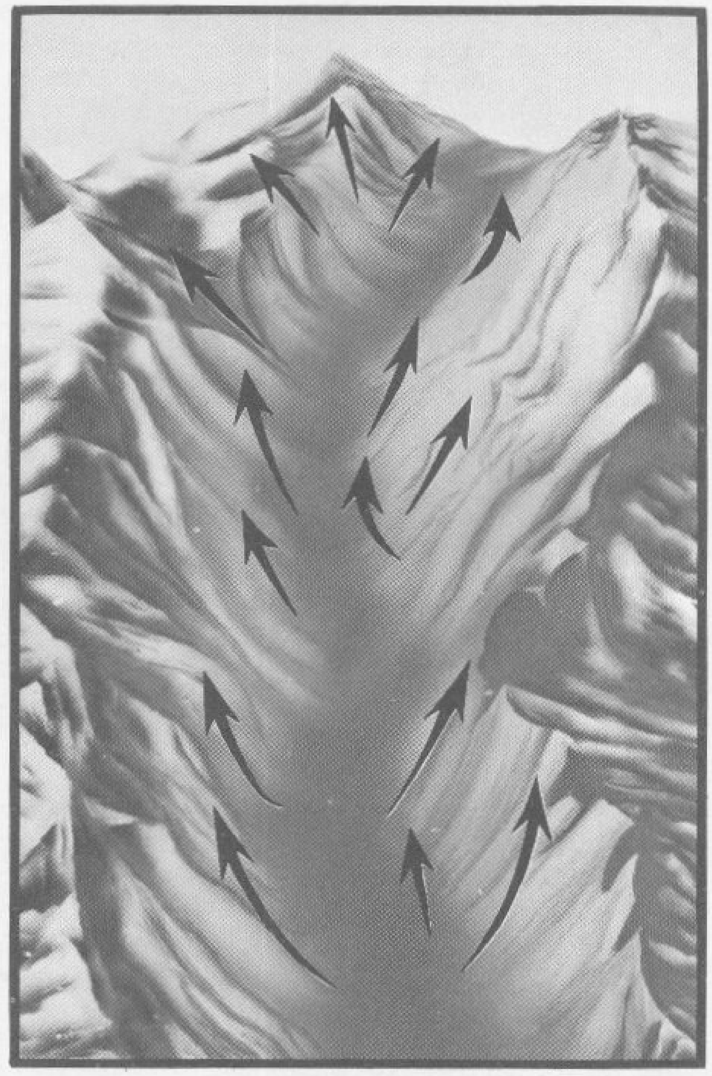
Valley winds and slope winds are not independent. A drainage valley or canyon bottom also has slope winds along its length, although perhaps not easy to identify separately. Proceeding upstream the combined flow continually divides at each tributary inlet in countless numbers of upravine and upslope components flowing off at the ridgetops. This outflow of air from a drainage system may account in considerable part for maintaining the reduced pressure in the upper reaches that causes the characteristic upvalley daytime wind above the surface.

Orientation of the topography is an important factor governing slope and valley wind strength and diurnal timing. Upslope winds begin as a gentle upflow soon after the sun strikes the slope. Upflow begins first on east slopes after daylight and increases in both extent and intensity as daytime heating continues. South and southwest slopes heat the most and have the strongest upslope winds. South slopes have their maximum windspeeds soon after midday and west slopes by about midafternoon.



Escape of warm air upward increases as upslope winds approach the crest.

Morning upslope winds flow straight up the slopes and minor draws.



Afternoon canyon winds turn upslope winds diagonally upcanyon.

Upslope windspeeds on south slopes are often several times those on the opposite north slopes.

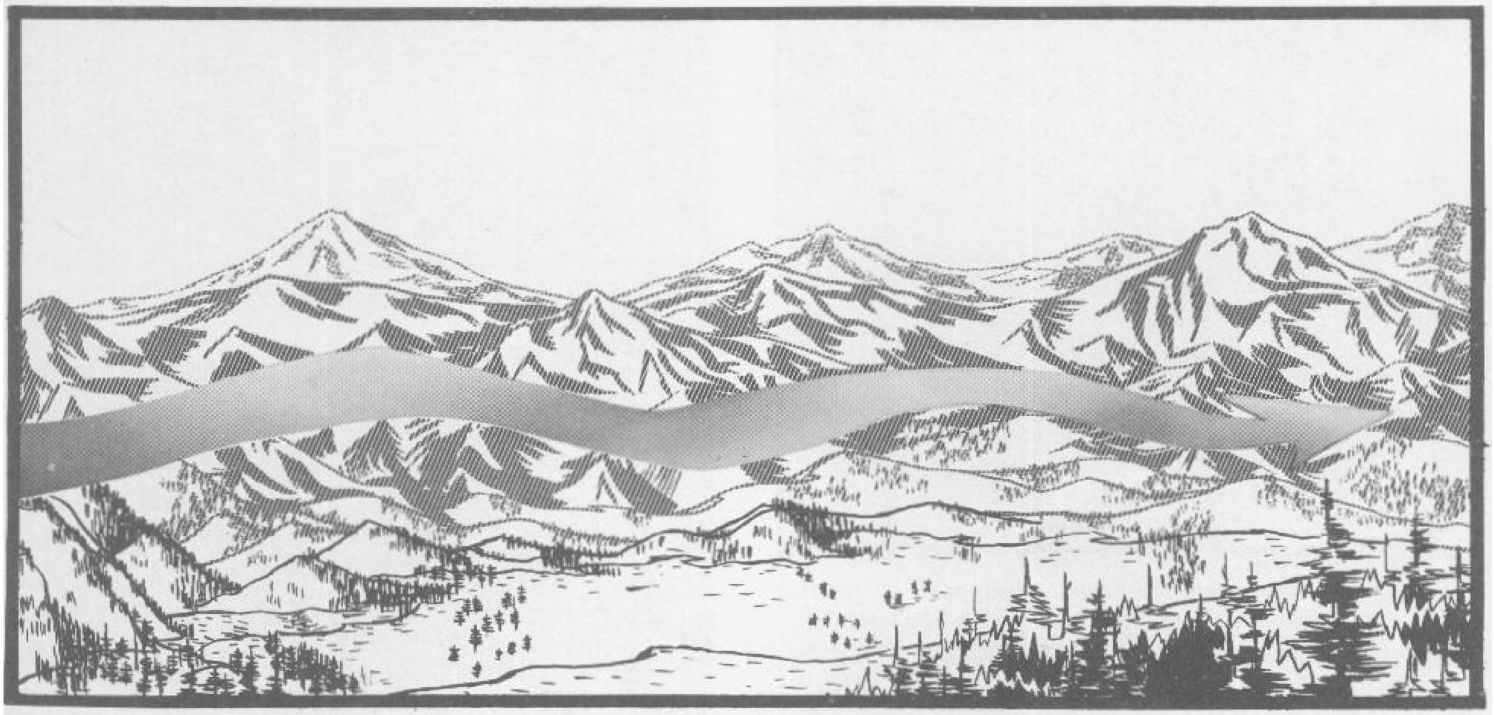
Upvalley winds begin later than the first slope winds. They may follow by an hour or more depending on size, configuration, and orientation of the drainage system. They reach maximum development later as well. In large drainages this is often midafternoon or later.

Strengthening of the whole valley wind system causes some change in upslope wind directions as the day advances. The first movement in the morning is directly up the slopes and minor draws toward the ridgetops. Then, as the valley wind picks up, it begins to swing the upslope winds in an upcanyon direction. By the time the valley wind reaches its maximum, the slope wind on the lower slopes, at least, may be turned completely in the upvalley direction.

The daytime valley wind often does not completely fill a valley or canyon, but flows along with its principal upper surface some distance below the ridgetops. In this situation the slope wind may maintain its early hour direction on the upper slopes throughout the day, while the direction on the lower slopes changes materially.

Strong upcanyon afternoon winds are often quite turbulent. Large eddies form in canyon bends and at tributary junctions, very much like those in upcanyon general windflow. These eddies tend to be more or less fixed in location, but pieces frequently break away and move along in the general upvalley flow. Air flowing in irregular channels the way may also pile up, then surge onward again.

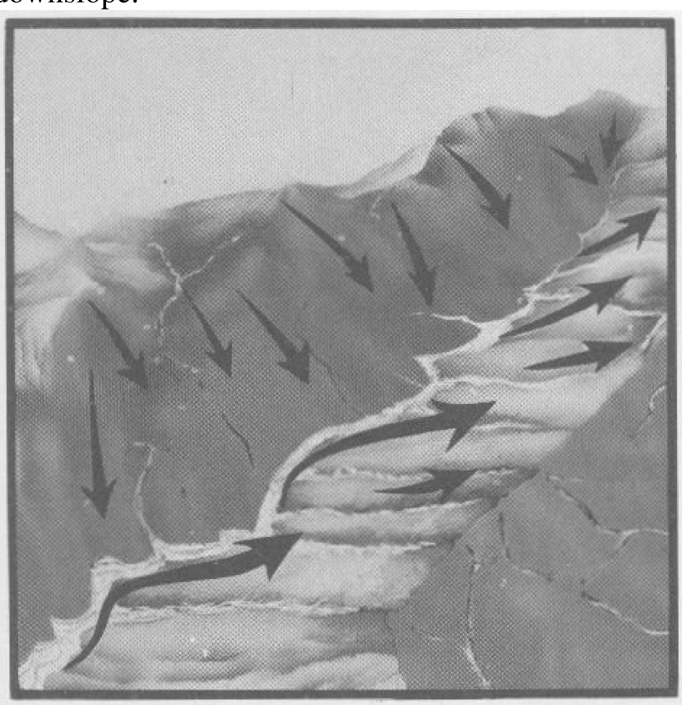
This turbulent, uneven upcanyon flow makes for many irregularities in the depth of upvalley winds. The observed behavior of forest fires indicates that the top surface of this flow may rise and sink in waves of varying frequency. Along the upper slopes it is not uncommon for fire spread to show alternate irregular intervals of upslope wind and upvalley



Undulating flow may occur with daytime canyon winds.

wind. Both wind speed and direction change materially and quickly with these fluctuations.

The transition from upslope to downslope wind begins soon after the first slopes go into afternoon shadow. Air heating on these ceases about this time and cooling then sets in. In the individual draws and on slopes going into shadow, the wind transition consists of dying of the upslope wind, a period of relative calm, and then gentle laminar flow downslope.

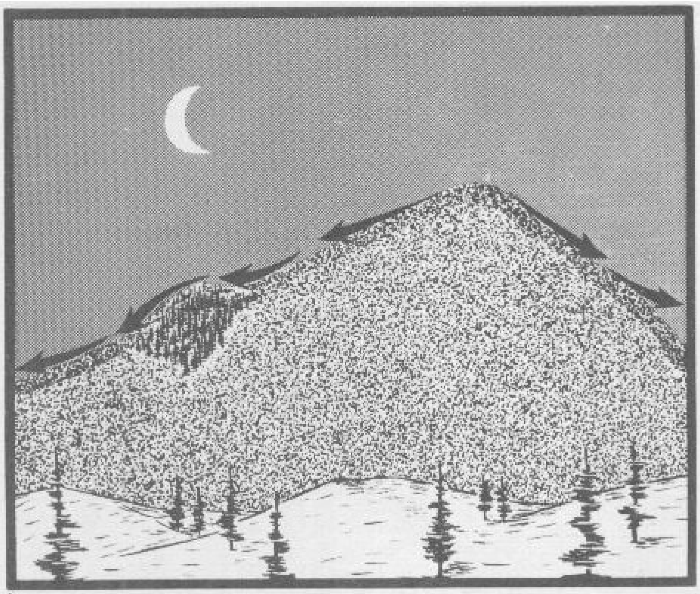


Valley and slope winds begin to shift direction as the first slopes go into shadow.

Where slopes with different aspects drain into a common basin, some slopes go into shadow before others and also before the general upflow in the area ceases. In many upland basins, the late afternoon surface winds are bent in the direction of the first downslope flow. They continue to shift as the downflow strengthens and additional slopes become shaded, until a 180-degree change in direction has taken place after all slopes are in shadow. This shift ordinarily takes place through the period required for cooling to set in on all exposures. Average windspeed may or may not decrease during the shift. In most well-drained upland areas it may approach zero locally, but with sustained windspeeds at many stations.

Downslope winds at night move in a shallow layer only a few feet deep except where there are obstructions to free flow. The most common obstructions are narrow crooked canyons and dense stand of tall timber. The flow is mostly laminar.

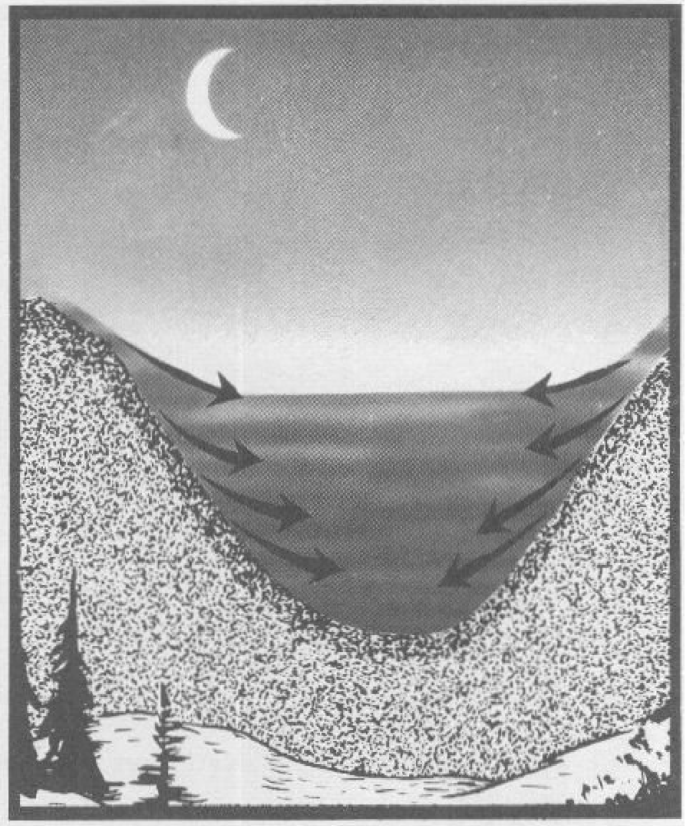
Cool air from the slopes accumulates in the natural drainage ways in the topography. It flows into low spots and overflows them when they are full, much like water. The principal force here is gravity. With weak to moderate temperature contracts, the resulting airflow tends to follow the steepest downward routes through the topography. Strong air temperature and density contrasts result in relatively higher airspeeds. With sufficient momentum the air then tends to flow in a straight path over minor topographic obstructions



Downslope winds are shallow.

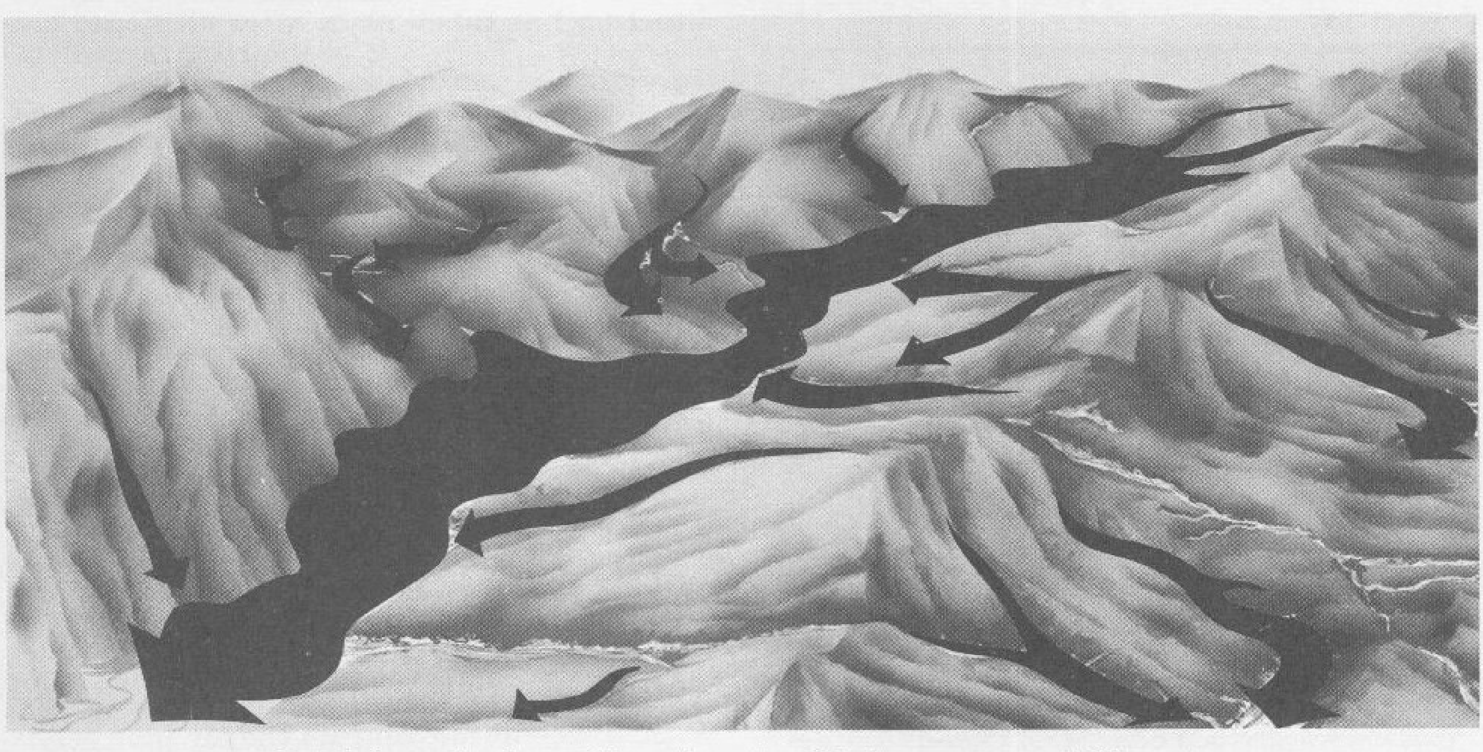
obstructions rather than turn to flow around them on its downward course.

Cool air accumulating in the bottom creates an inversion which increases in depth and strength during the nighttime hours. The top of the inversion may often attain maximum height sometime after midnight, but cooling at the surface continues as a rule throughout the night. The downslope winds from above continue downward until they reach air of their own temperature. There, they fan out horizontally over the canyon or valley. This may be at or near the top of the inversion or some distance below it.



Air from downslope winds contributes to a deepening inversion.

The accumulation of cooling downslope winds in the canyons at night increases the pressure in the upper drainage areas, and causes the whole valley air system to move outward toward adjoining plains. In many localities the beginning of the general down-



Cool air from the slopes follows the natural drainage ways much like water.

valley flow varies from night to night, often by as much as 2 hours, probably as a result of both pressure and temperature variations. Downslope winds below the inversion frequently slacken by midnight or somewhat later, while the downvalley flow proper may continue until just before the beginning of the morning return upflow. The speeds of both downslope and downvalley winds are much lower than daytime upvalley and upslope winds.

Interaction Of Slope And Valley Winds And Gradient Winds

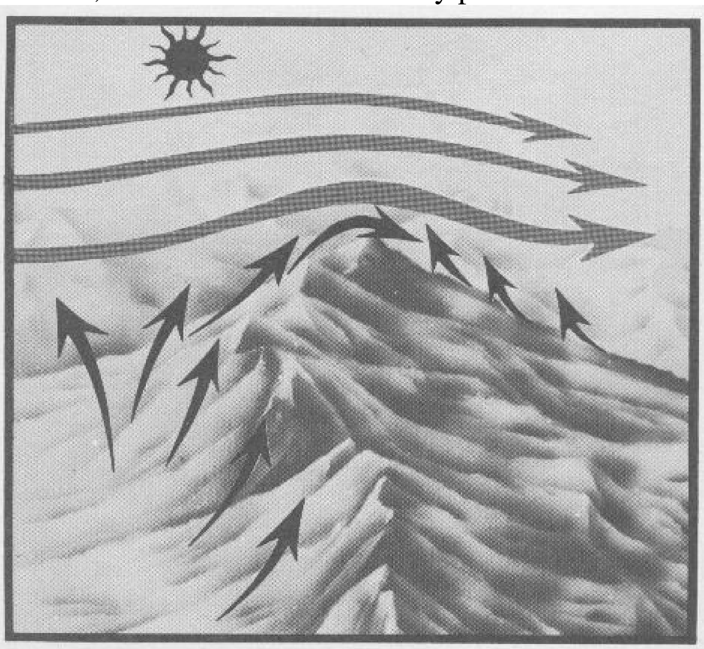
Slope and valley wind systems are subject to interruption at any time by the gradient winds. But even when this does not occur, differences in heating over the terrain itself may set up local pressure differences and induce changes in local wind direction from that expected. For example, it is not uncommon for daytime downslope and downcanyon winds to occur in one part of a drainage system and upslope and upcanyon in another. Sometimes reversal from upflow to downflow may occur by midday or soon after, as has been noted in both the Sierra Nevada and southern California mountains.

Summer midday upslope winds in mountain topography tend to hold weak winds aloft above the ridgetops. Very frequently these daytime upper winds are felt only in the highest peaks. In this situation, surface winds are virtually pure convection winds. Upslope winds usually dominate the ridges and saddles, with upslope and upvalley winds combining to define wind speeds and directions at the lower elevations.

Late afternoon weakening of upslope winds and the onset of downslope flow lowers the gradient wind level back onto exposed slopes and ridgetops. If this wind is relatively cool and not too strong, some portion of it near the surface may join with the downslope wind, considerably increasing its force. Proceeding downslope, the combined flow may retain the shallow characteristics of a downslope wind or, if sufficient in volume, may fill whole canyons and then take on the characteristics of an accelerated downvalley wind.

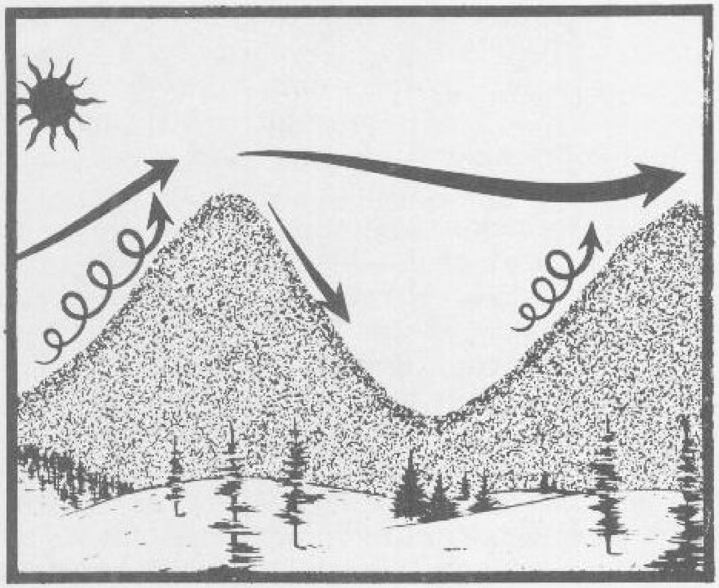
A westerly gradient wind across a north-south or easterly drainage may be kept aloft by continued heating on a western slope until an inversion has formed on the le side to seal off a valley system. Then, the upper wind may maintain a nearly horizontal trajectory above the valley level, resulting in a two-storied surface wind structure during the evening and nighttime hours. Below the inversion, winds are cool, downslope and downcanyon and predominately gentle. Above the inversion, wind on upper slopes and ridges is generally warmer, stronger, and more turbulent.

Gradient winds of cold or relatively dense air tend to follow the surface of the topography as noted in the case of foehn winds. When these blow up- or down-valley, they frequently blow along the surface,

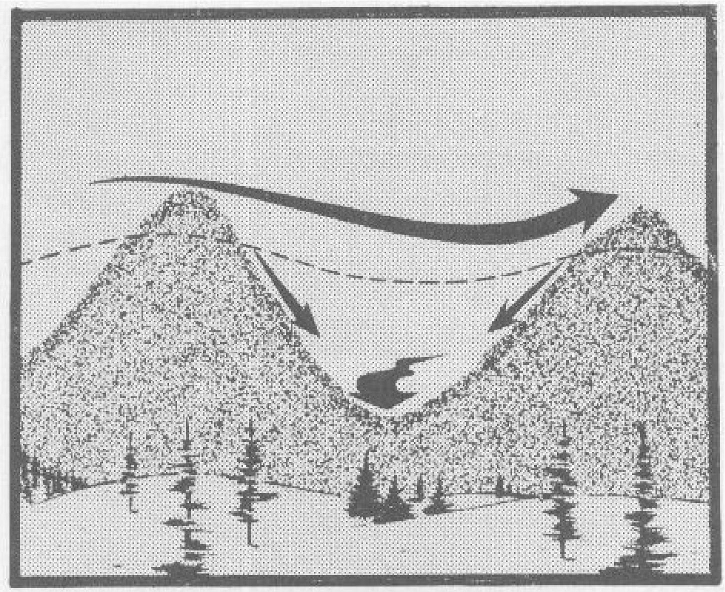


Convective winds hold gradient winds aloft under strong heating.

Gradient winds drop down onto exposed peaks and ridges at night.



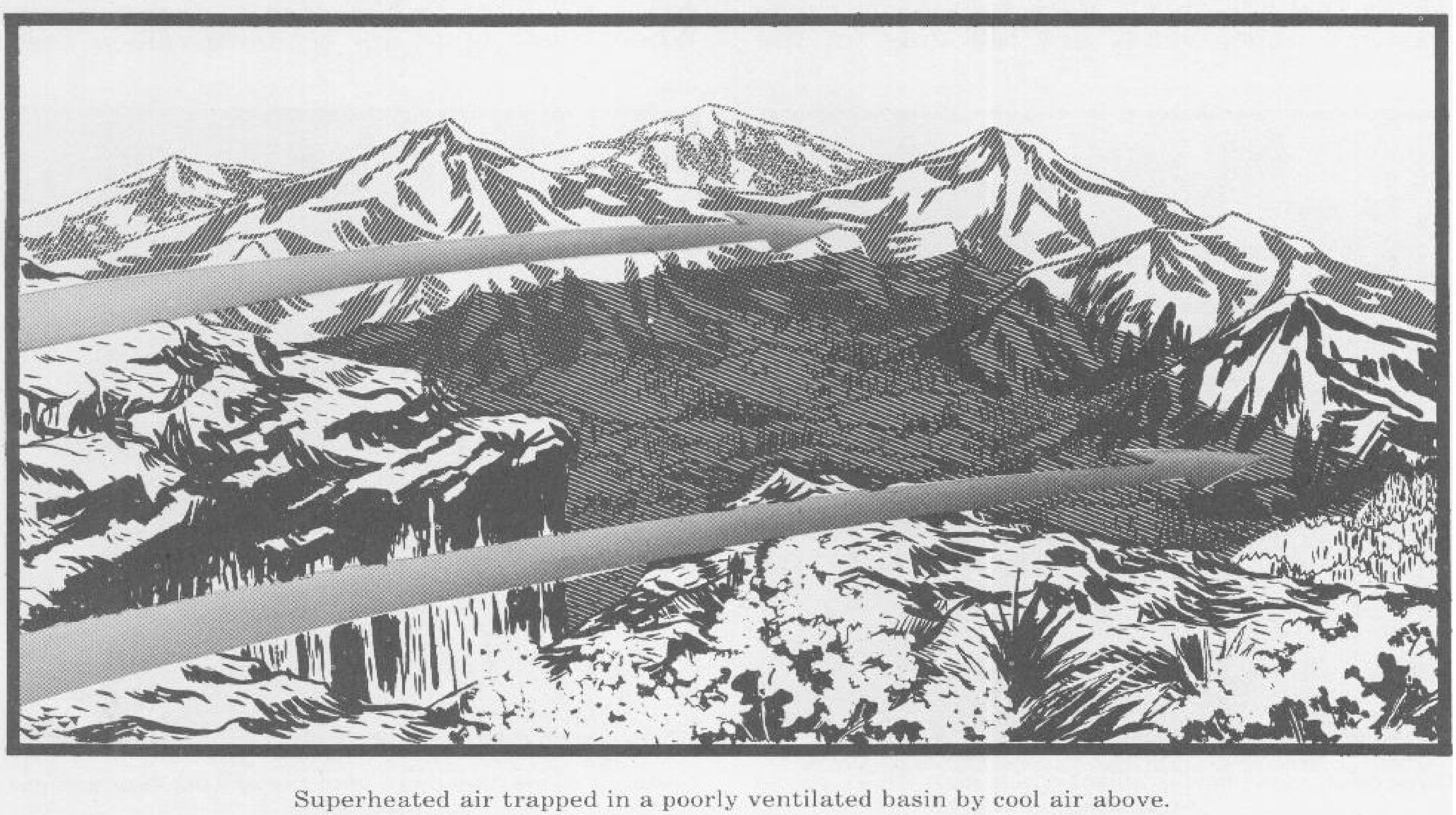
Late afternoon heating can hold westerly winds aloft until inversions form in the lee valleys.



Slope and valley winds below an inversion are undisturbed by gradient winds above.

minimizing any surface heating effects. They also blow up and down slopes in crossing wide valleys. If reasonably strong, however, momentum of the windflow may carry it across many narrow mountain canyons without significantly disturbing the flow in canyon.

Cool and gently moving winds aloft sometimes prevent slope and valley winds from developing in poorly ventilated canyons and basins, even with strong heating. This can occur most readily in local areas surrounded by sharp ridges of relatively uniform height and which drain through narrow canyons or gorges. Nighttime inversions form in the basin, and in the cooling process the air aloft settles down onto the upper slopes. If the upper air is dense, heating the following day, even though intense, may not lift it back up above the ridgetops. This creates a very strong superadiabatic lapse rate and a potentially explosive situation in the trapped air below.



Superheated air trapped in a poorly ventilated basin by cool air above.

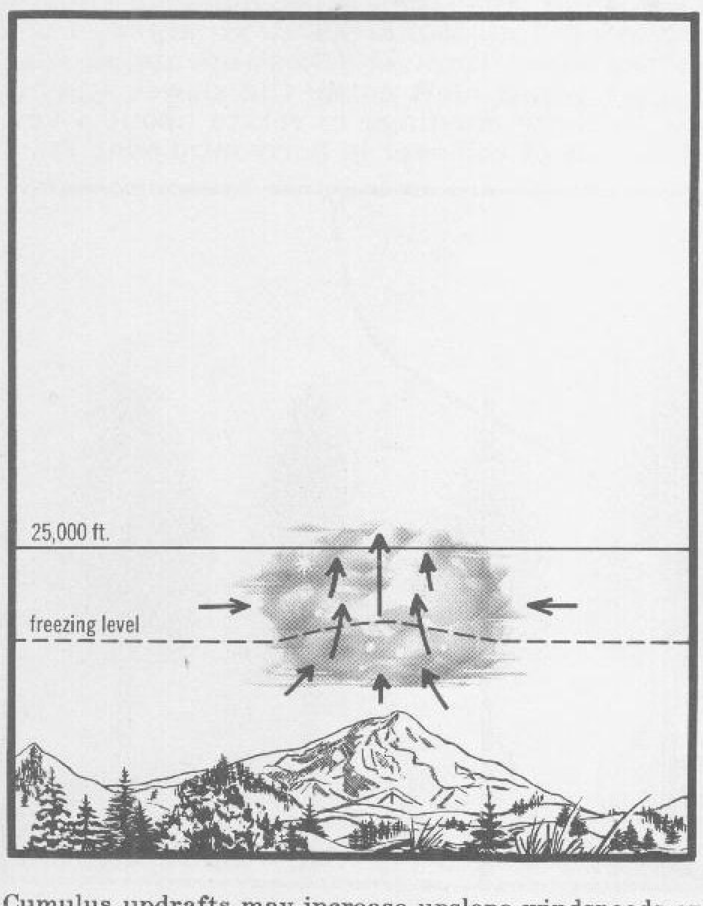
Thunderstorm Phenomena

Three special types of local winds are associated with cumulus cloud growth and thunderstorm activity: updrafts predominating in growing cumulus clouds, downdrafts in the later stages of full thunderstorm development, and cooled air following the storm, possibly generating squall characteristics.

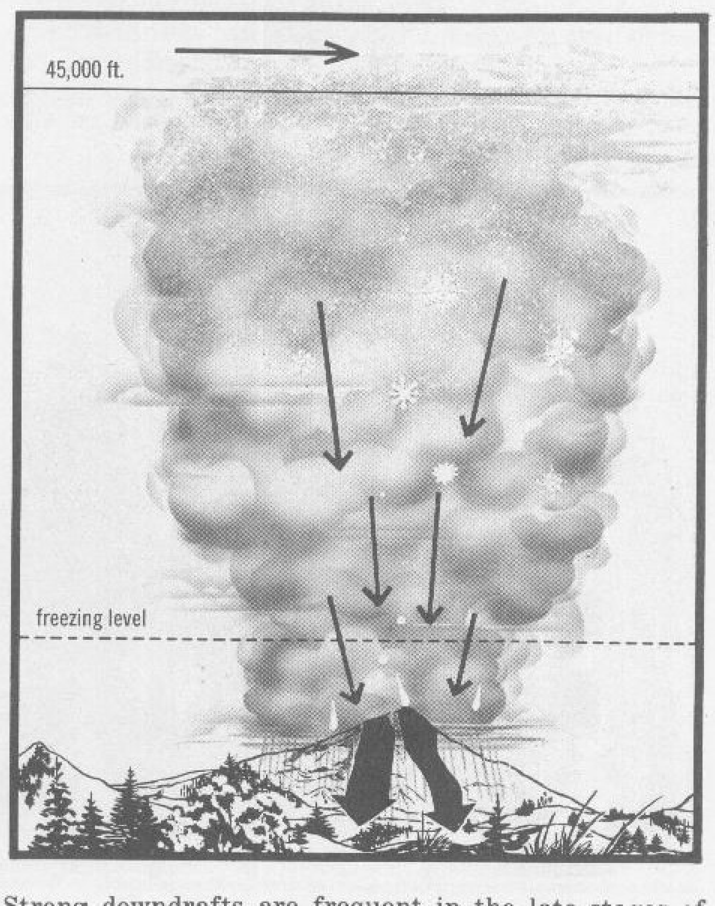
There are always strong updrafts within growing cumulus clouds, often 30 m.p.h. or more. Ordinarily, the warm air fed into the cloud base is drawn from a large pool of surrounding air. The indraft to the cloud base in this case is not felt very far below or away from the cloud cell. Cells that form over peaks and ridges as described earlier, however, may actually increase the speeds of the upslope winds that initiated the cloud formation. A cumulus cloud formed elsewhere that drifts over one of these ridge plumes may increase the upslope winds in similar fashion while the cloud grows with renewed vigor. With continuing drift, the cloud may draw the ridgetop convection with it for a considerable distance.

Fully developed thunderstorm cells contain a large volume of relatively cold air before they finally dissipate. After the thunderstorm cell has passed its most active stage, this cold air may cascade to the surface as a strong downdraft. In level terrain this becomes a surface wind guided by direction of the general wind and favorable airflow channels. In mountainous terrain it continues its downward path into the principal drainage ways. Speeds of 20 to 30 m.p.h. are common. This gives the air sufficient momentum to traverse at least short adverse slopes in this downward plunge. These speeds and surface roughness also cause the wind to be gusty. Although these winds strike suddenly and violently when they occur they are of only short duration, usually a few minutes.

Thunderstorms in the mountainous West often cool sizable masses of air over areas of 100 to several hundred square miles. Occurring as they do in the warm summer months, these cool air masses are in strong temperature contrast with their surroundings. As this air spreads out and settles to the lower levels, squall winds develop ahead of the leading edge or front. These are strong and gusty, lasting but a few minutes. They behave much like the winds in squall lines ahead of cold fronts, but are on a smaller geographic scale. They begin and end quickly, but may travel out many miles beyond the original storm area.



Cumulus updrafts may increase upslope windspeeds on the higher ridges.



Strong downdrafts are frequent in the late stages of cumulonimbus activity.

Surface Winds In Forests

Forest vegetation is part of the friction surface which determines how the wind blows near the ground. Forests are characteristically rough surfaces and thus contribute to air turbulence, eddies and the like. They also have the distinction of being more or less porous, allowing some air movement through as well as over and around the vegetation.

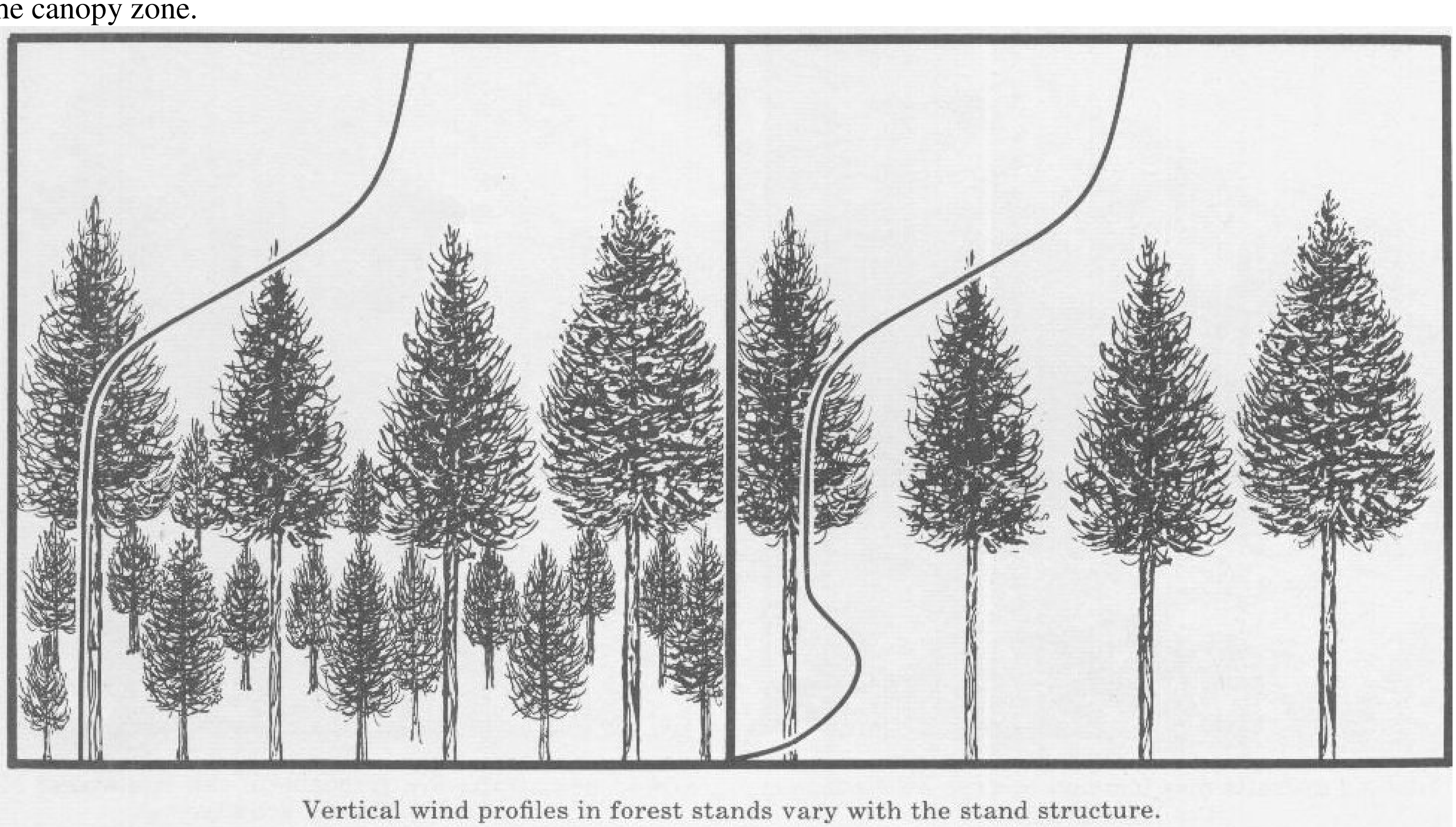
Average windspeeds over open, level ground decrease quite rapidly in the last 20 feet above the ground, reaching zero windspeed at the surface. Where the surface is covered with low-growing, dense vegetation like grass or brush, it is satisfactory for most weather purposes to consider the effective friction surface as the average height of the vegetation, disregarding the air flowing through it. In areas forested with trees, however, airflow within and below the tree canopies is often appreciable.

The leaf canopy in a forest is very effective in slowing down wind movement because of its large friction area. In forests of shade-tolerant species where the canopy extends to near ground level or in stands with understory vegetation, windspeed is nearly constant from the surface up to ear the tops of the crowns. Above the crowns, windspeed increases much as it does over level ground. In forest stands that are open beneath the main tree canopy, air speed increases above the surface to the middle of the space below the crowns and then decreases again in the canopy zone.

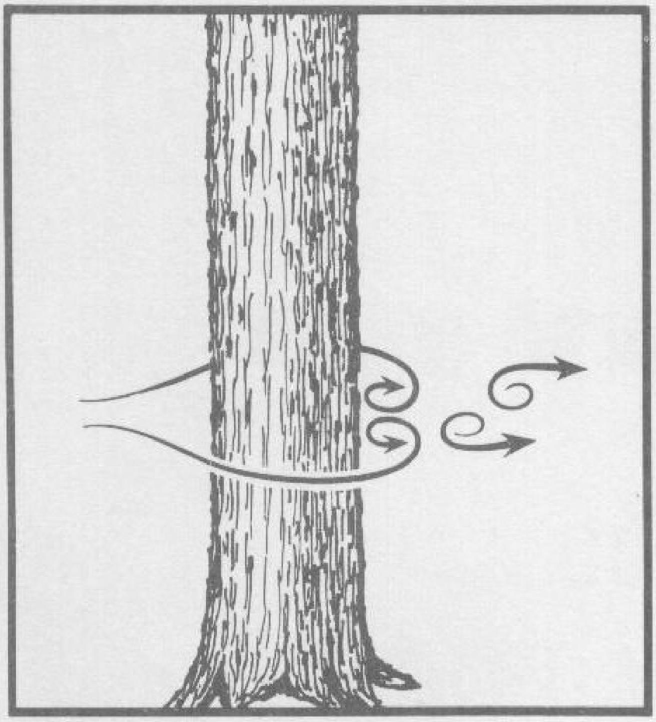
How much the windspeed is reduced inside the forest depends on its detailed structure and on windspeed above the forest canopy or as measured out in the open away from the forest. The drag of any friction surface is relatively much greater at high windspeeds than it is with low speeds. At low windspeeds the forest may reduce the speed of the wind blowing through it only slightly. For example, a 4-m.p.h. wind measured in the open might be slowed to 2.5 m.p.h. at the same height inside the forest. At high windspeeds in the open the greater drag keeps the wind in the forest still at low speed. Thus, a 30-m.p.h. wind might be reduced to 4 or 5 m.p.h. in the forest.

The quantitative relationships cited might apply to an 80-feet-tall stand of second-growth pine with normal stocking. They would vary considerably, however, between different species and types of forest. Deciduous forests have a further seasonal variation. Trees bare of leaves still have a significant effect in limiting surface windspeeds, though far less than when in full leaf.

Local eddies are common in forest stands. One of the most frequently observed is found in the lee of each tree stem. Surface fires running along the forest floor are caught in these and thus burn most hotly n te lee sides of the trunks. On a larger scale, eddies often form in forest openings. The higher winds



Vertical wind profiles in forest stands vary with the stand structure.



Forest fires running along the surface burn intensely in these eddies.

aloft cause the slower moving air in these openings to rotate about a vertical axis or roll over in horizontal eddy fashion. The surface wind direction is then frequently directly opposite to the direction above the treetops.

The edges of tree stands often cause roll eddies to form similar to those associated with bluffs. Wind blowing against the stand often produces small transient eddies on the windward side, while those in the lee of a forest are mostly larger, and more fixed in location with subeddies breaking off and moving downwind.

Strong surface heating, as on warm sunny days, adds to the complexity of these forest airflow patterns. Thermal turbulence is added to the generally turbulent flow through open timber stands as it is to the flow above a closed forest canopy. The flow beneath a dense canopy is affected but little except where holes let the sun strike bare ground or litter on the forest floor. These become hotspots over which there is a general upwelling of warm air through the canopy. This rising air is replaced by gentle inflow from surrounding shaded areas. Thermal turbulence on the lee side of a forest stand may often be enough to disguise or breakup any roll eddies that tend to form.

Air beneath the forest canopy under a nighttime inversion is usually calm. An important exception is a nighttime downdraft on slope or valley bottom in which there is a forest with open space beneath the main tree canopy. Here the flow is confined mostly to the open space with calm prevailing in the canopy region. Forests with no free space beneath the main tree canopy are effective barriers to downslope winds. Here the flow is diverted around the stand or confined to stream channels, roadways, or other openings that cut through it.