

CONVECTIVE WINDS

Winds of local origin—convective winds caused by local temperature differences—can be as important in fire behavior as the winds produced by the synoptic-scale pressure pattern. In many areas they are the predominant winds in that they overshadow the general winds. If their interactions are understood, and their patterns known, the changes in behavior of wildfires can be predicted with reasonable accuracy. Fires occurring along a coastline will react to the changes in the land and sea breezes. Those burning in mountain valleys will be influenced by the locally produced valley and slope winds. Certainly there will be times when the convective winds will be severely altered or completely obliterated by a strong general wind flow. These cases, in which the influences of the general winds on fire behavior will predominate, must be recognized.

Convective Winds

In the absence of strong synoptic-scale pressure gradients, local circulation in the atmosphere is often dominated by winds resulting from small-scale pressure gradients produced by temperature differences within the locality. Air made buoyant by warming at the surface is forced aloft; air which is cooled tends to sink. Buoyant air is caused to rise by horizontal airflow resulting from the temperature-induced small-scale pressure gradients.

In different convective circulation systems, either the vertical or the horizontal flow may be the more important but both are part of the same system. Hence, convective winds here refer to all winds—up, down, or horizontal—that have their principal origin in local temperature differences. This is somewhat different from common meteorological usage, wherein convection implies upward motion only.

Convective winds may be augmented, opposed, or eliminated by airflow having its origin in the larger pressure systems. The influence of these general winds on the convective wind systems varies with the strength of the general wind, its direction relative to the convective circulation, and the stability of the lower atmosphere.

The nature and strength of convective winds vary with many other 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, cloud cover, nature of the terrain and its cover such as water, vegetation, or bare ground, and the moisture and temperature structure of the overlying atmosphere.

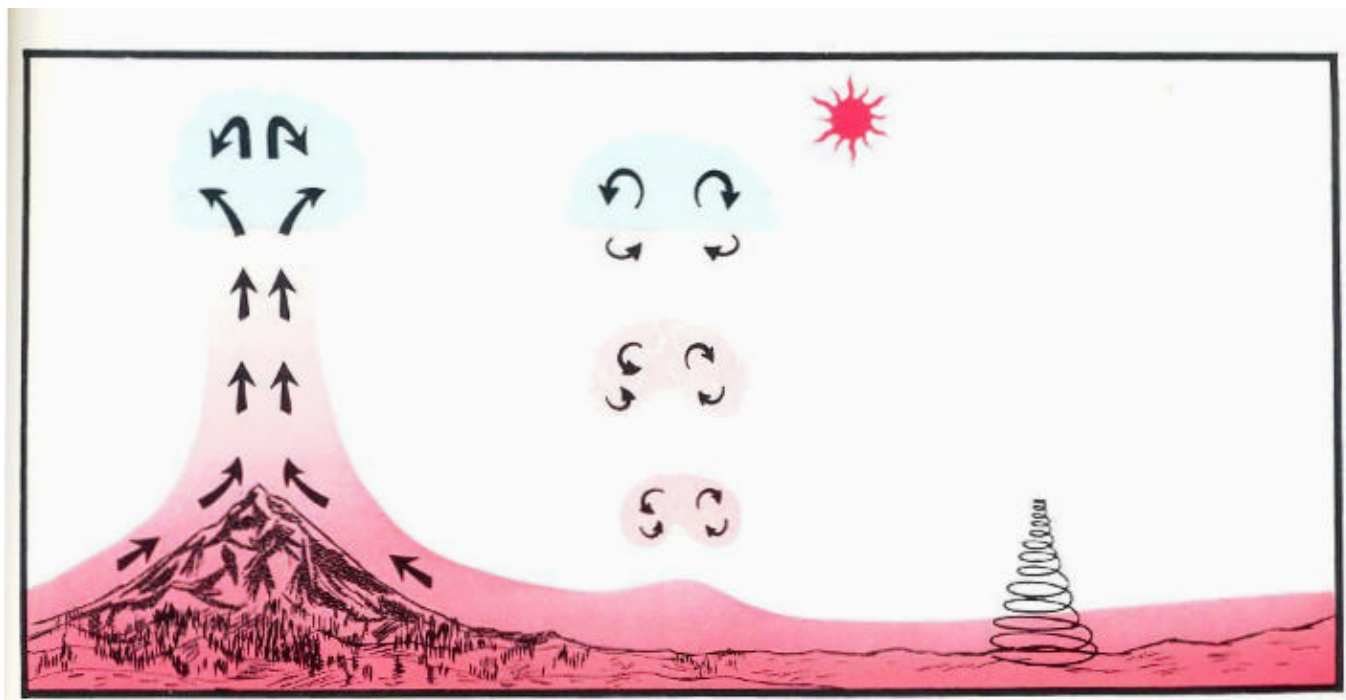
The strong temperature dependence of convective winds make local temperature observations useful indicators of probable wind behavior. Simultaneous measurements may show significant horizontal temperature gradients. In the absence of upper-air soundings,

mountaintop and valley-bottom readings give fair approximations of the temperature lapse rate and associated stability or instability. Height of the nighttime inversion may usually be located in mountain valleys by traversing side slopes and by taking thermometer readings.

Strong surface heating produces the most varied and complex convective wind systems. Warmed air adjacent to heated slopes tends to be forced upslope to the crest where it flows off in a more-or-less continuous stream. These convective currents frequently cause daytime cumulus clouds so often observed over mountain peaks and ridges. In generally flat terrain, air heated at the surface tends to remain in stagnant layers because of inertia, until it reaches a critical point of instability or is released by mechanical triggering. The escaping air usually takes the form of intermittent bubbles that break free and are forced aloft by surrounding denser air. As they ascend, the bubbles grow by expansion and by mixing with surrounding air. These, too, may form cumulus clouds. Superheated air may escape also in the form of upward-spiraling whirlwinds or dust devils. These vortexes draw on new supplies of heated air as they move along the surface.

Air that is cooled near the surface almost invariably flows downward along the steepest route available, seeking the lowest levels. En-route, if it should meet colder air beneath it, the downflowing air spreads out on top of the colder layer.

Other types of local convective circulations involving both vertical and horizontal movement occur where there are differences in heating between sizeable adjacent areas. Most familiar among these are the land and sea breezes found along ocean shores and around the larger inland lakes and bays.



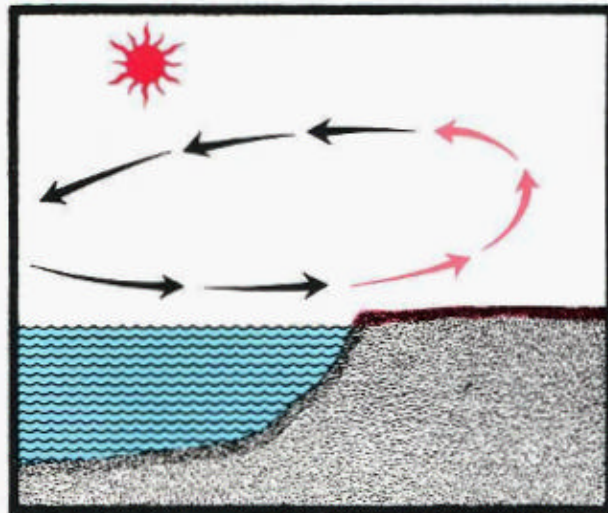
Strong surface heating produces several kinds of convective systems. Upslope winds develop along heated slopes. Superheated air in flat terrain escapes upward in bubbles or in the form of whirlwinds or dust devils.

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 that over the nearby water. In chapter 2 we considered in some detail the several reasons why land surfaces become warmer than water surfaces during the daytime. As a result of this local-scale pressure difference, a sea **breeze begins** to flow inland from over the water, forcing the warm air over the land to rise and cool adiabatically. In the absence of strong general winds, this air flows seaward aloft to replace air which has settled and moved toward shore, and thus completes the circulation cell.

The surface sea breeze begins around mid-forenoon, strengthens during the day, and ends around sunset, although the times can vary of local considerably because conditions of cloudiness and the general winds. The breeze begins at the coast, then gradually pushes farther and farther inland during the day, reaching its maximum penetration about the time of maximum temperature.

The **land breeze** at night is the reverse of the daytime sea-breeze circulation. At night, land surfaces cool more quickly than water surfaces



As land surfaces become warmer than adjacent water surfaces during the daytime, and air over the land becomes warm and buoyant, the denser sea breeze begins to flow inland from over the water and forces the less-dense air upward.

(discussed in chapter 2). Air in contact with the land then becomes cooler than air over adjacent water. The increase in air density causes pressure to become relatively higher over the land than over the water, and this pressure difference, in turn, causes air to flow from the land to the water. The air must be replaced, but any return flow aloft is likely to be so weak and diffuse that it is lost in the prevailing general winds.

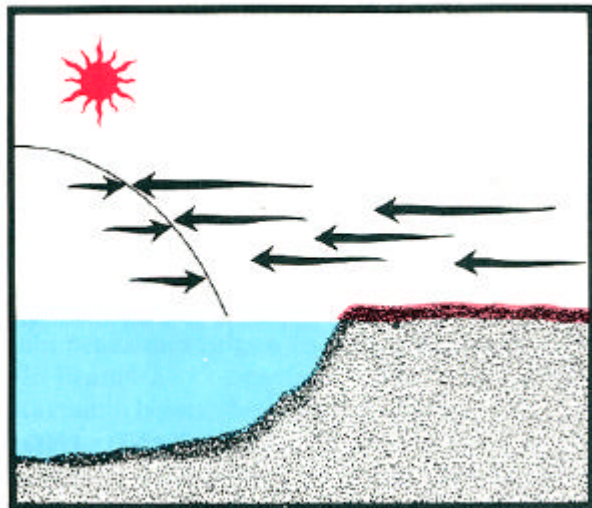
The land breeze begins 2 to 3 hours after sunset and ends shortly after sunrise. It is a more gentle flow than the sea breeze, usually about 3 to 5 miles per hour. The land air, having been cooled from below by contact with the ground, is stable. The land breeze is, therefore, more laminar and shallower than the sea breeze.

The daily land and sea breezes tend to occur quite regularly when there is no significant influence from the general wind flow. When general winds are sufficiently strong, however, they usually mask the land and sea breezes. A general wind blowing toward the sea opposes the sea breeze and, if strong enough, may prevent its development. In any case the sea breeze is delayed. Depending on the strength of the general wind, this delay may extend into the afternoon. This often results in a “piling up” of marine air off the coast. Then, when the local pressure difference becomes great

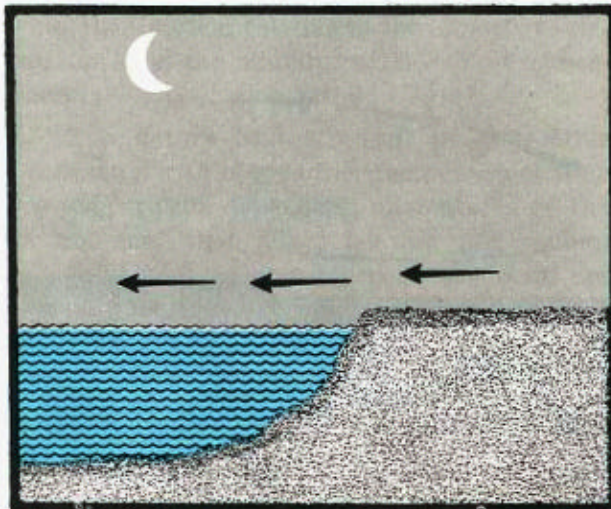
enough, this sea air moves in-land with the characteristics of a small-scale cold front. Air behind the front is initially cool and moist but warms rapidly as it moves over sun-warmed land.

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

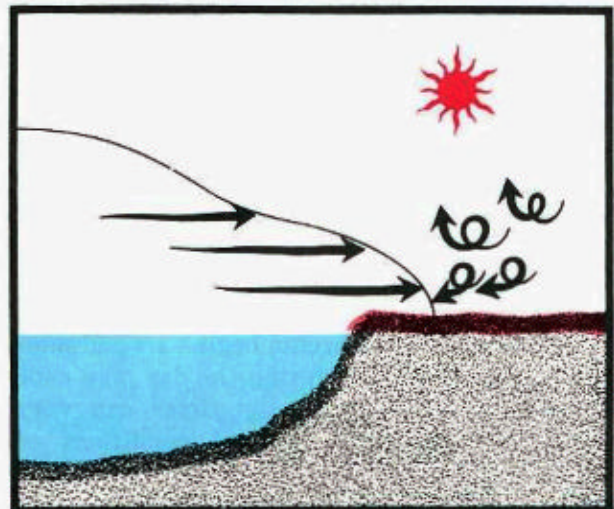
General winds, either in the direction of the land or sea breeze, or parallel to the coast,



A general wind blowing toward the sea operates against the sea breeze and, if strong enough, may block the sea breeze entirely.



At night, land surfaces cool more quickly than water surfaces. Air in contact with the land becomes cool and flows out over the water as a land breeze, displacing the warmer air.



If marine air has been piled up over the water by an offshore wind, it may rush inland like a small-scale cold front when the local pressure difference becomes great enough.

tend to mask the true land- or sea-breeze component. Strong general winds produce mechanical mixing which tends to lessen the temperature difference between the land and the sea surfaces. Thus the sea-breeze component becomes weak and only slightly alters the general wind flow. General winds also tend to mask out the closed-cell feature of the land- and sea-breeze circulations by overshadowing the return flow aloft. With an onshore general wind aloft, for example, there is no return flow in the daytime sea-breeze circulation.

General winds along an irregular or crooked coastline may oppose a land or sea breeze in one sector and support it in another. Oftentimes, too, shifting general winds may cause periodic reversals of these effects in nearby localities, and may result in highly variable local wind patterns.

Land and sea breezes occur along much of the Pacific coast, the Gulf of Mexico, and the Atlantic seaboard. Eastern and western land and sea breezes differ in their respective behaviors due to marked differences in general circulation patterns, temperature contrasts, and topography. Whether or not these factors are significant locally depends on the local climate



General winds along an irregular or crooked coastline may oppose a sea breeze in one sector and not in another.

and on the shape and orientation of the shoreline and inland topography.

Gulf and Atlantic Breezes

In the East, land and sea breezes are most pronounced in late spring and early summer, when land and water temperature differences are greatest, and they taper off toward the end of the warm season as temperature differences decrease. They are sufficiently strong during the spring and fall fire seasons to warrant consideration as important fire-weather elements in coastal areas.

Land- and sea-breeze circulations in the East are more often dominated by changes in the general wind pattern than they are in the West. Otherwise, the eastern land and sea breeze represents a more simple situation than the western because coastal topography is flat and uniform.

During the fire season in the East, general circulation patterns are such that on both the Gulf and Atlantic shores there are frequent periods of onshore or offshore winds strong enough to block or mask out land- and sea-breeze development. Onshore general winds almost always mask sea-breeze effects. During periods of gentle to moderate offshore winds, on the other hand, the sea breeze may develop and move inland. Against an opposing general wind, however, the sea breeze moves forward behind a small-scale cold front. This moves slowly, perhaps 3 or 4 miles an hour, and at times may oscillate back and forth with the varying force of the general wind. In addition to the rapid changes in wind speed and direction associated with a cold-front passage, a small area may thus be subjected to several of these passages over a considerable time. At this slow and intermittent pace, the sea breeze may have penetrated inland only a few miles by late afternoon.

Another feature of this type of sea breeze is that it is operating in an area of convergence. This is conducive to turbulent vertical motion in addition to the above-mentioned horizontal surface disturbances. **This combination can create critical fire-weather situations,** particularly in view of the fact that this type of sea breeze is prone to occur on high fire-danger days.

The reverse land breeze often becomes just part of the offshore general wind and thereby loses its identity.

Pacific Coast Sea Breeze

The Pacific coastal area sea breeze is at its peak at the height of the summer fire season. It is an important feature of the summer weather along much of the Pacific coast. Water temperatures there are much lower than along the Gulf of Mexico and the Atlantic coast. Intense daytime land heating under clear skies is an additional factor in producing greater land-water temperature differences along the Pacific coast. The sea breeze is, therefore, stronger along the western than the eastern coasts. It is a daily summertime occurrence along the Pacific coast except on rare occasions when it is opposed by the general circulation.

Normally the general wind serves to strengthen the Pacific coast sea breeze. During the summer months, the semipermanent North Pacific High is located in the general area between Hawaii and Alaska. Flow from this high to the California Low results in onshore surface winds along most of the Pacific coast. This seasonal flow, called the Pacific coast monsoon, begins in spring and lasts until fall. The sea breeze is superimposed on the monsoon circulation. During the day, air from the ocean moves inland, rises as it is heated, mixes with the upper winds, and is replaced on the seaward side by gradually settling air from the general circulation.

Since the monsoon flows onshore both day and night, it tends to weaken, or reduce to a negligible amount, the night land breeze. However, this opposition of forces also slows down the onshore monsoon at night. During the day, the sea breeze, assisted by the monsoon, brings in a fresh surge of marine air. Because of this assistance, the marine layer is thicker, and moves farther inland, than does the sea breeze in the East.

The Pacific sea breeze brings relatively cool, moist marine air to the coastal areas. The passage of the leading edge of this air—the sea-breeze front—is marked by a wind shift and an increase in wind speed. Often it is accompanied by fog or low stratus clouds, particularly in the morning hours. Within the first few

miles inland, however, the marine air is subjected to heating as it passes over the warmer land. If the marine-air layer is shallower than normal, this air may soon become almost as warm as the air it is replacing. The strong temperature contrasts then remain near the coast while the warmed sea breeze may penetrate many miles beyond.

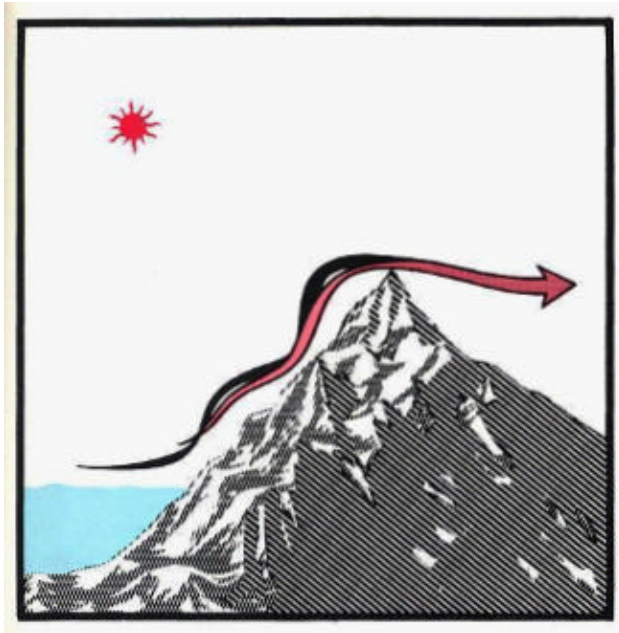
Thus the effect of the sea breeze on fire behavior can vary considerably. Where the marine air is not modified appreciably, its lower temperatures and higher humidities produce less dangerous fire weather. Where the marine air is modified extensively by heating, the temperature and humidity changes with the sea-breeze front become negligible, while the shifting wind direction and increase in wind speed and gustiness can be a serious detriment to fire control.

Because of surface friction, the sea breeze often moves inland more rapidly at the top of the marine layer than at the surface. Instability and convective mixing caused by surface warming then tend to bring the sea breeze aloft down to the surface, so that the sea-breeze front appears to progress on the surface in jumps or surges. The motion is somewhat analogous to that of the forward portions of the endless metal tracks on a moving tractor.

The Pacific sea breeze is characterized by considerable thermal turbulence and may extend inland 30 to 40 miles or more from the water under favorable conditions. The depth of the sea breeze is usually around 1,200 to 1,500 feet, but sometimes reaches 3,000 feet or more. Its intensity will vary with the water-land temperature contrast, but usually its speed is around 10 to 15 miles per hour.

Mountains along the Pacific coastline act as barriers to the free flow of surface air between the water and the land. On seaward-facing slopes the sea breeze may combine with upslope winds during the daytime, thus transporting modified marine air to the higher elevations in the coastal mountains.

River systems and other deep passes that penetrate the coast ranges provide the principal inland sea-breeze flow routes. The flow of cool, moist air is sufficient to carry tremendous amounts of marine air inland, helping to maintain inland summer humidities at moderate levels in the areas opposite the passes. Here, the



Mountains along the coastline act as barriers to the free flow of air between the water and the land. On seaward-facing slopes, the sea breeze may combine with upslope winds during the daytime and bring modified marine air to higher elevations.

sea breeze joins with afternoon upvalley and upcanyon winds, resulting in a cooler, relatively strong flow. In broad valleys, this flow takes on the usual sea-breeze characteristics, but in narrow canyons and gorges it may be strong and very gusty as a result of both mechanical and thermal turbulence.

The 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



River systems and other deep passes that cut through coast ranges provide the principal sea-breeze flow routes.

downvalley and downcanyon flow is, like the normal land breeze, a relatively shallow and low-speed wind system.

Small-scale diurnal circulations similar in principle to land and sea breezes occur along the shores of 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. On a summer afternoon it is not unusual for most shore stations to experience onshore winds.

SLOPE AND VALLEY WINDS

Winds in mountain topography are extremely complex. Part of the time, the general winds associated with larger scale pressure systems dominate the surface layer. But when larger scale pressure systems weaken, the general winds lessen. Then, in the presence of strong daytime heating or nighttime cooling, convective winds of local origin become important features of mountain weather. These conditions are typical of clear summer weather in

which there is a large diurnal range of surface air temperatures.

General and convective winds may displace, reinforce, or oppose each other. Their relationship to each other can change quickly—often with surprising rapidity. Variations between different terrain features—sometimes separated only by yards—are often noted. The convective activity may dominate the observed surface wind in one instance, and in another it

may permit the speed and direction of winds aloft to dominate the surface flow through the mixing process.

The interactions between airflow of different origins, local pressure gradients caused by nonuniform heating of mountain slopes, and the exceedingly complex physical shapes of mountain systems combine to prevent the rigid application of rules of thumb to convective winds in mountain areas. Every local situation must be interpreted in terms of its unique qualities. Wind behavior described in this section is considered typical, but it is subject to interruption or change at virtually any time or place.

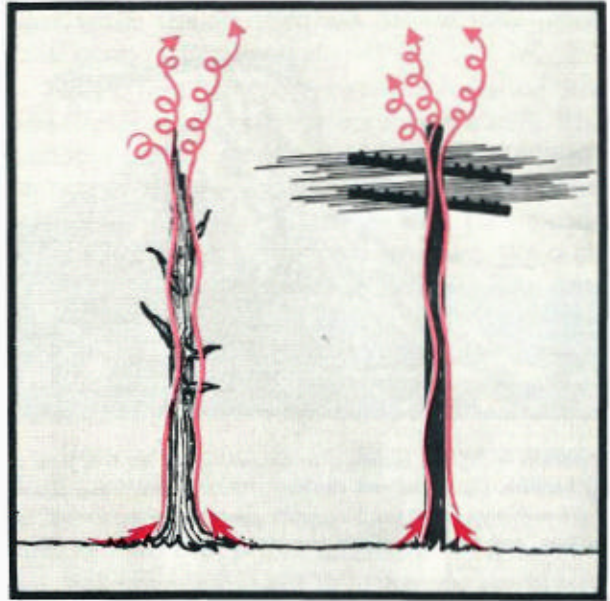
Differences in air heating over mountain slopes, canyon bottoms, valleys, and adjacent plains result in several different but related wind systems. These systems combine in most instances and operate together. Their common denominator is upvalley, upcanyon, upslope flow in the daytime and downflow at night. They result from horizontal pressure differences, local changes in stability that aid vertical motion, or from a combination of the two.

Slope Winds

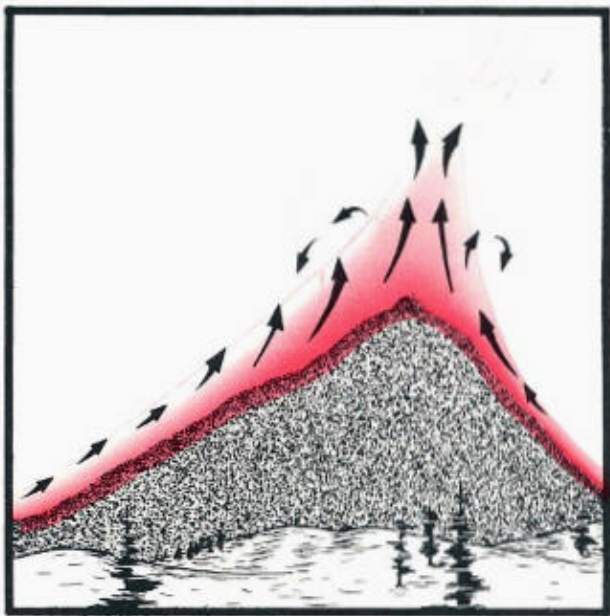
Slope winds are local diurnal winds present on all sloping surfaces. They flow upslope during the day as the result of surface heating, and downslope at night because of surface cooling. Slope winds are produced by the local pressure gradient caused by the difference in temperature between air near the slope and air at the same elevation away from the slope.

During the daytime the warm air sheath next to the slope serves as a natural chimney and provides a path of least resistance for the upward flow of warm air. Ravines or draws facing the sun are particularly effective chimneys because of the large area of heated surface and steeper slopes; winds are frequently stronger here than on intervening spur ridges or uniform slopes. Upslope winds are quite shallow, but their depth increases from the lower portion of the slope to the upper portion. Turbulence and depth of the unstable layer increase to the crest of the slope, which is the main exit for the warm air. Here, momentum of the upflowing air, convergence of upslope winds from opposite slopes, and mechanical turbulence combine to make the ridge a very turbulent region where much of the warm air escapes aloft.

The crests of higher ridges are also likely to experience the influence of the general wind flow, if that flow is moderate or strong.



Air heated by contact with vertical or sloping surfaces is forced upward and establishes natural chimneys through which warm air flows up from the surface.



Upslope winds are shallow near the base of slopes but increase in depth and speed as more heated air is funneled along the slope. Warm air bubbles forced upward cause turbulence which increases the depth of the warmed layer.

At night the cool air near the surface flows downslope much like water, following the natural drainage ways in the topography. The transition from upslope to downslope wind begins soon after the first slopes go into afternoon shadow and cooling of the surface begins. In individual draws and on slopes going into shadow, the transition period consists of (1) dying of the upslope wind, (2) a period of relative calm, and then (3) gentle laminar flow downslope.

Downslope winds are very shallow and of a slower speed than upslope winds. The cooled denser air is stable and the downslope flow, therefore, tends to be laminar.

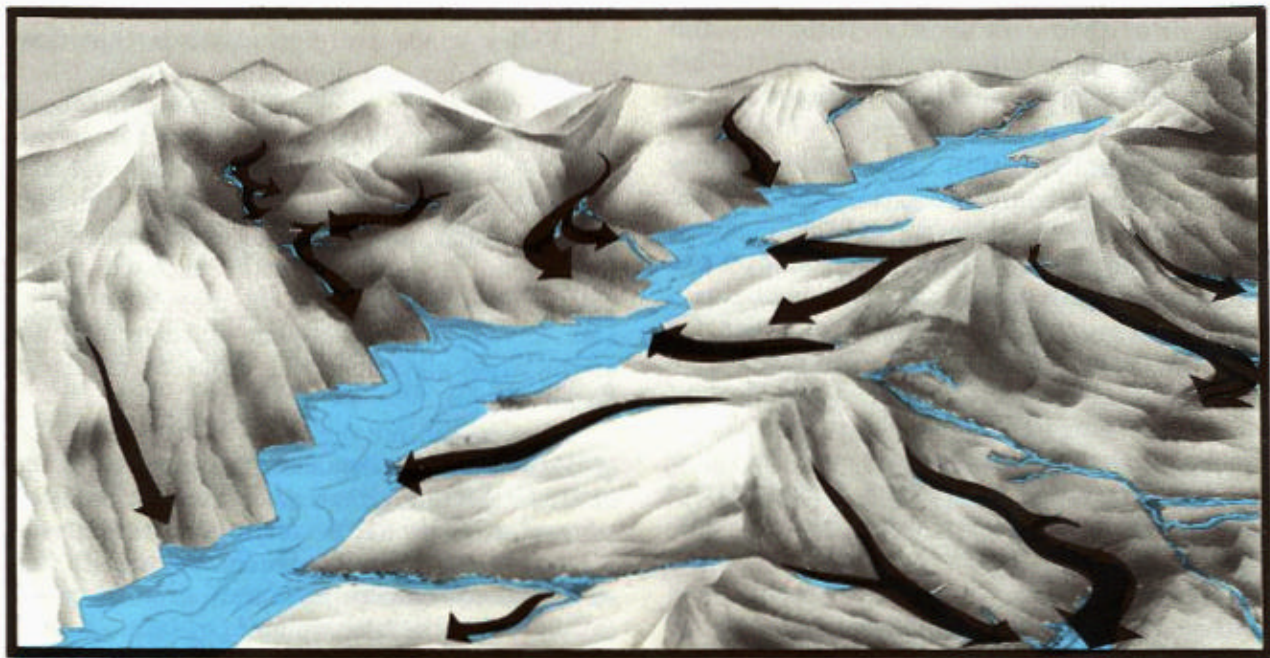
Downslope winds may be dammed temporarily where there are obstructions to free flow, such as crooked canyons and dense brush or timber. Cool air from slopes accumulates in low spots and overflows them when they are full. The principal force here is gravity. With weak to moderate temperature contrasts, the airflow tends to follow the steepest downward routes through the topography. Strong air temperature contrasts result in relatively higher air speeds. With sufficient momentum, the air tends to flow in a



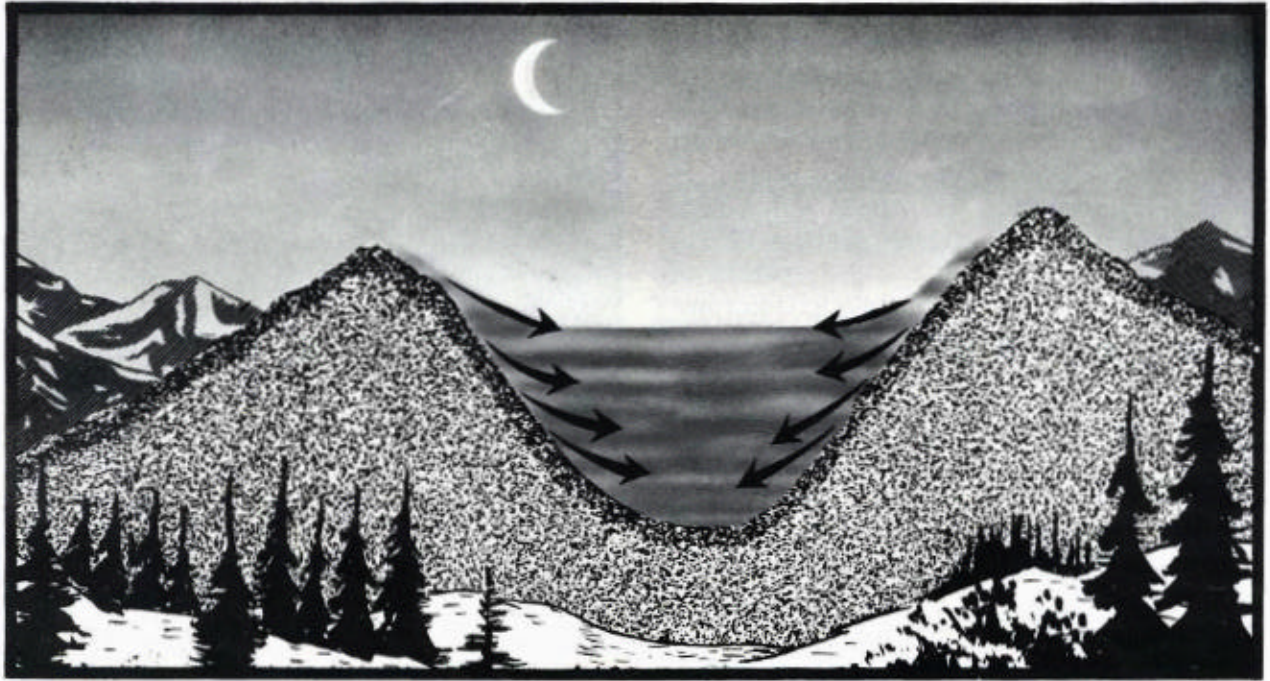
Downslope winds are shallow, and the flow tends to be laminar. The cold air may be dammed by obstructions such as dense brush or timber.

straight path over minor topographic obstructions rather than to separate and flow around them on its downward course.

Cool, dense air accumulates in the bottom of canyons and valleys, creating an inversion which increases in depth and strength during the night hours. Downslope winds from above



At night, the cool denser air near the surface of slopes flows downward, much like water, following the natural drainage ways in the topography. The principal force is gravity.



Cool, dense air settles in the bottom of canyons and valleys, creating an inversion which increases in depth and strength during the night hours. Cool air flows outward over the valley bottom when it reaches air of its own density.

the inversion continue downward until they reach air of their own density. There they fan out horizontally over the canyon or valley. This may be either near the top of the inversion or some distance below the top.

Theoretically, both upslope and downslope winds may result in a cross-valley circulation. Air cooled along the slopes at night flows downward and may be replaced by air from over the valley bottom. Air flowing upslope in the daytime may be replaced by settling cooler air over the center of the valley. The circulation system may be completed if the upward flowing air, on reaching the upper slopes, has cooled enough adiabatically to flow out over the valley and replace air that has settled. During strong daytime heating, however, cross-valley circulation may be absent. Upflowing air is continually warmed along the slopes. Adiabatic cooling may not be sufficient to offset the warming, and the warmer air is forced aloft above the ridgetops by denser surface air brought in by the upvalley winds.

Valley Winds

Valley winds are diurnal winds that flow **upvalley** by day and **downvalley** at night. They are the result of local pressure gradients caused by differences in temperature between air in the valley and air at the same elevation over the adjacent plain or larger valley. This temperature difference, and the resulting pressure difference and airflow, reverses from day to night. During the day, the air in the mountain valleys and canyons tends to become warmer than air at the same elevation over adjacent plains or larger valleys.

One reason for the more intensive heating of the mountain valley air is the smaller volume of air in the valley than over the same horizontal surface area of the plain. The rest of the volume is taken up by landmass beneath the slopes. A valley may have only from one-half to three-fourths the volume of air as that above the same horizontal surface area of the plain.

Another reason is the fact that the mountain

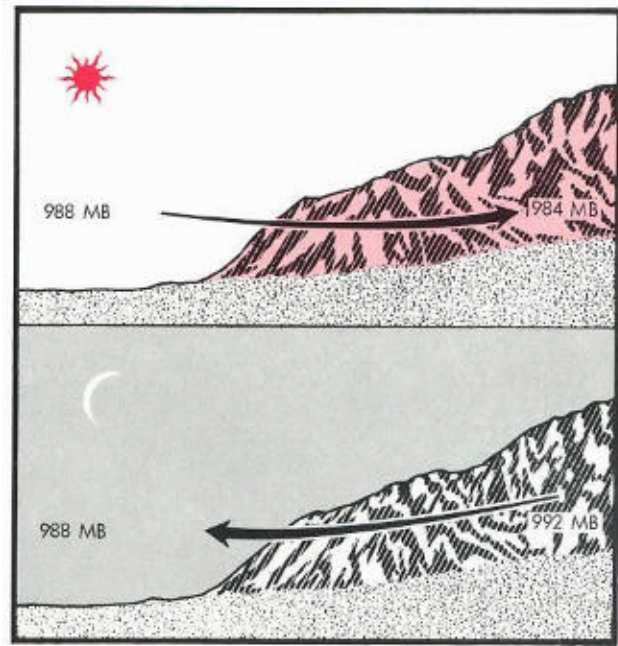
valley air is somewhat protected by the surrounding ridges from the general wind flow. The valley air is heated by contact with the slopes, and the resulting slope-wind circulation is effective in distributing the heat through the entire mass of valley air. As the valley air becomes warmer and less dense than the air over the plain, a local pressure gradient is established from the plain to the valley, and an upvalley wind begins.

Whereas upslope winds begin within minutes after the sun strikes the slope, the up-valley wind does not start until the whole mass of air within the valley becomes warmed. Usually this is middle or late forenoon, depending largely on the size of the valley. The upvalley wind reaches its maximum speed in early afternoon and continues into the evening. Upvalley wind speeds in larger valleys are ordinarily from 10 to 15 m.p.h. The depth of the upvalley wind over the center of the valley is usually about the same as the average ridge height.

Strong upvalley and upcanyon winds may be quite turbulent because of the unstable air and the roughness of the terrain. Eddies may form at canyon bends and at tributary junctions. Along upper ridges particularly, the flow tends to be quite erratic. Wind speed and direction may change quickly, thus drastically affecting fire behavior.

Slopes along the valley sides begin to cool in late afternoon and, shortly after they come into shadow, cool air starts flowing downslope. Cool air accumulates in the valley bottom as more air from above comes in contact with the slopes and is cooled. Pressure builds up in the valley, causing the upvalley wind to cease. With continued cooling, the surface pressure within the valley becomes higher than the pressure at the same elevation over the plain, and a downvalley flow begins.

The transition from upvalley to downvalley flow takes place in the early night—the time depending on the size of the valley or canyon and on factors favoring cooling and the establishment of a temperature differential. The transition takes place gradually. First, a down-slope wind develops along the valley floor, deepens during the early



Upvalley and downvalley winds are the result of local pressure gradients caused by differences in temperature between air in the valley and air at the same elevation over the adjacent plain. The flow reverses from day to night.

night, and becomes the downvalley wind. The downvalley wind may be thought of as the exodus or release of the dense air pool created by cooling along the slopes. It is somewhat shallower than the up-valley wind, with little or no turbulence because of the stable temperature structure of the air. Its speed is ordinarily somewhat less than the upvalley wind, but there are exceptions in which the downvalley wind may be quite strong.

The downvalley wind continues through the night and diminishes after sunrise.

Valley winds and slope winds are not independent. A sloping valley or canyon bottom also has slope winds along its length, although these winds may not be easy to distinguish from valley winds. Proceeding upstream during the daytime, the combined flow continually divides at each tributary inlet into many up-ravine and upslope components to the ridge-tops. As the valley-wind system strengthens during the day, the direction of the upslope wind is affected. The first movement in the morning is directly up the slopes and minor draws to the ridgetop. Then, as the speed of the valley wind picks up, the upslope winds are changed to a more upvalley direction. By the



Th. combined upslope and upvalley winds proceed upstream during the day and continually divide at each tributary into many upravine and upslope components.

time the valley wind reaches its maximum, the slope winds, on the lower slopes at least, may be completely dominated by the upvalley wind. Along the upper slopes, the direction may continue to be upslope, because the upvalley wind does not always completely fill the valley.

Nighttime downslope winds are similarly affected. When the downvalley wind is fully developed, it dominates the flow along the slopes, particularly the lower portion, so that the observed wind direction is downvalley.

Effects of Orientation and Vegetation

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. Therefore, they begin first on east-facing slopes after daybreak and increase in both intensity and extent as daytime heating continues. South and southwest slopes heat the most and have the strongest upslope winds. South slopes reach their maximum wind speeds soon after midday, and west slopes by about midafternoon. Upslope wind speeds on south slopes may be several times greater than those on the opposite north slopes.



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

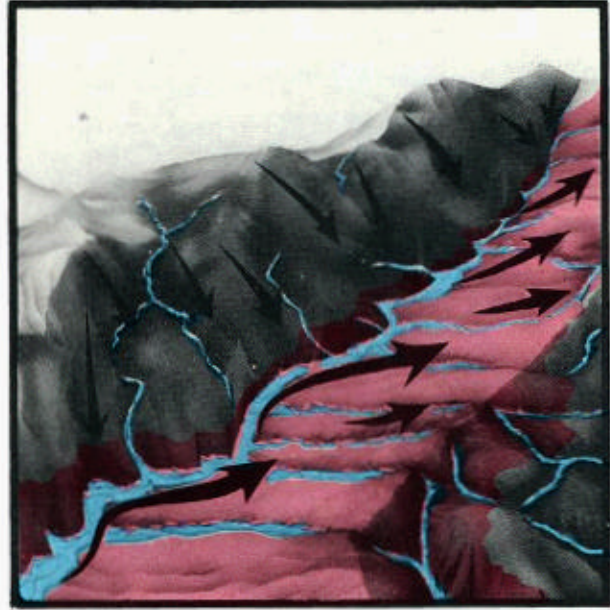
Where slopes with different aspects drain into a common basin, some slopes go into shadow before others and also before the upvalley wind ceases. In many upland basins, the late afternoon upvalley winds are bent in the direction of the first downslope flow. They continue to shift as the downslope flow strengthens and additional slopes become shaded, until a 180-degree change in direction has taken place some time after sunset.

The vegetative cover on slopes will also affect slope winds and, in turn, valley winds. Bare slopes and grassy slopes will heat up more readily than slopes covered with brush or trees. Upslope winds will therefore be lighter on the brush- or tree-covered slopes. In fact, on densely forested slopes the upslope wind may move above the treetops, while at the surface



As the upvalley wind picks up during the day, the upslope winds are turned to a more upvalley direction.

there may be a very shallow downslope flow because of the shade provided by the canopy.



Downslope winds begin as soon as slopes go into shadow. Late afternoon upvalley winds are turned in the direction of the first downslope flow.

Downslope winds at night on densely forested slopes are affected by the presence or absence of a dense understory. Where there is an open space between the tree canopy and the surface, the downslope flow will be confined to the trunk region while calm prevails in the canopy region. A forest with a dense understory is an effective barrier to downslope winds. Here, the flow is diverted around dense areas, or confined to stream channels, roadways, or other openings cut through the forest.

INTERACTION OF VALLEY AND SLOPE WINDS WITH GENERAL WINDS

Slope and valley wind systems are subject to interruption or modification at any time by the general winds or by larger scale convective wind systems.

Midday upslope winds in mountain topography tend to force weak general winds aloft above the ridgetop. The general wind flow goes

over the rising currents above the ridge. These rising currents may be effective in producing or modifying waves in the general wind flow. Frequently, the daytime upper winds are felt only on the highest peaks. In this situation, the surface winds, except on the highest peaks, are virtually pure convective winds. Upslope winds



General winds are modified by local wind flow. Weak general winds may exist only at or above ridgetops when strong upslope winds predominate. Upslope winds may establish or intensify wave motion in the general wind flow.



Late afternoon weakening of upslope winds and the onset of downslope flow in the early evening allow the general winds to lower onto exposed upper slopes and ridgetops. If the air being brought in by the general wind is relatively cold, this wind may add to the downslope wind on the lee side of ridges and result in increased speed.

dominate the saddles and lower ridges and combine with upvalley winds to determine wind speeds and directions at the lower elevations.

Late afternoon weakening of upslope winds and the onset of downslope flow in the early evening allow the general winds to lower onto the exposed upper slopes and ridgetops. In the Far West, air in the flow aloft from the North Pacific High is subsiding and, therefore, commonly warm and dry. At night, this air may be found at higher levels at least as far inland as the Sierra-Cascade Range. A fire burning to a ridgetop under the influence of upslope afternoon winds may flare up, and its spread may be strongly affected as it comes under the influence of the general wind flow. Similar phenomena may occur in mountainous country elsewhere.

Valley winds are affected by the general wind flow according to their relative strengths, directions, and temperatures. The degree of interaction also varies from day to night.

The general wind has its maximum effect on valley winds during the daytime when a strong general wind blows parallel to the valley. If the general wind is

blowing in the direction of the upvalley wind and the air is relatively unstable, the influence of the general wind will be felt down to the valley floor. The resulting surface wind will be a combination of the general wind and the upvalley wind. When the general wind blows in the direction opposite to the upvalley wind, it extends its influence some distance down into the valley and the observed surface wind will be the resultant of the upvalley and general winds.

General winds blowing at right angles to the axis of a valley during the daytime have much less influence on the valley wind pattern than those blowing along the valley. The ridges tend to shield the valley circulation from the effects of the general wind.

The relative coldness or density of air being brought in by the general winds is an important factor. Relatively warm air will continue to flow aloft without dropping into valleys and canyons and disturbing the convective wind systems. But cold, relatively dense air combined with strong general wind flow tends to follow the surface of the topography, scouring out valleys and canyons and completely erasing the valley wind systems. Such effects are common in cold air following the passage of a cold

front, and in deep layers of cold marine air along the Pacific coast. In these situations the general wind flow is dominant.

These effects are most pronounced when the general wind flow is parallel to the axis of the valley. Strongwinds blowing across narrow valleys and canyons may not be able to drop down into them since momentum may carry the airflow across too quickly. Then, too, there are in-between situations where the general wind flow only partly disturbs the valley wind systems. General winds warm adiabatically as they descend the slopes on the windward side of a valley. If the descending air reaches a temperature equal to that of the valley air, it will leave the slope and cross the valley. The cooler the air flowing in with the general wind, the farther it will descend into the valley.

General winds at night usually have much less effect on valley wind systems than during the daytime. Ordinarily a nighttime inversion forms in the valleys, and this effectively shields the downvalley wind from the general wind flow. Again, there are important exceptions that must be considered.

If the air being brought in by the general wind flow is relatively cold and the direction is appropriate, the general wind can combine with downslope and downvalley winds and produce fairly strong surface winds, particularly during the evening hours. Later during the night, however, further cooling will usually establish a surface inversion and the general wind influence will be lifted to the top of the inversion.

Another important exception is the action of lee-side mountain waves. As was mentioned in the previous chapter, when mountain waves extend down to the surface they will completely obscure valley wind systems. In foehn wind situations this may occur during the day or night, but after the first day of the foehn wind, it is most common during the evening hours.

Downslope Afternoon Winds

An exception to the normal upcanyon, upslope, daytime flow occurs frequently enough on the east slopes of the Pacific Coast Ranges to



A nighttime inversion in a valley effectively shields the downslope and downvalley winds from the general wind flow above.

warrant further discussion. During the forenoon, in the absence of an overriding general wind flow, local winds tend to be upslope and flow up the draws on both the west and east sides of the Coast Ranges. Usually, the flow through gaps and saddles is easterly because of the stronger heating on the east side in the forenoon. The two flows meet in a convergence zone on the west side of the ridge. By midday the flow up the west slopes has increased, most likely because of the sea breeze or a strengthening of the monsoon circulation due to intensification of the thermal trough. The convergence zone has moved eastward across the ridge, and the flow through the gaps has changed to westerly.

Waves form in this westerly flow, which first remain aloft on the lee side of the mountains, and later surface to cause strong down-slope winds on the east side. Downslope afternoon winds are commonly three times as strong as the forenoon upslope winds. In some areas, downslope afternoon winds occur nearly every day during the warm season, while in other areas they occur only occasionally. The time of the wind shift from upslope to downslope on the east side may vary from late forenoon to late afternoon, but most frequently it is around noon or early afternoon. On some days, up-slope winds redevelop in late afternoon as the mountain waves go aloft. On other days, the downslope afternoon winds diminish and change



During the forenoon in the western Coast Ranges, local winds tend to be upslope and upcanyon on both the east and west sides. The two flows meet in a convergence zone on the west side. If a westerly flow aloft develops, it temporarily rides over the convergence zone and the easterly upslope winds.



On some afternoons the convergence zone moves east as the westerly flow increases. If waves with suitable length and amplitude form in the flow aloft, strong winds blow down the east slopes and east-facing canyons.

WHIRLWINDS

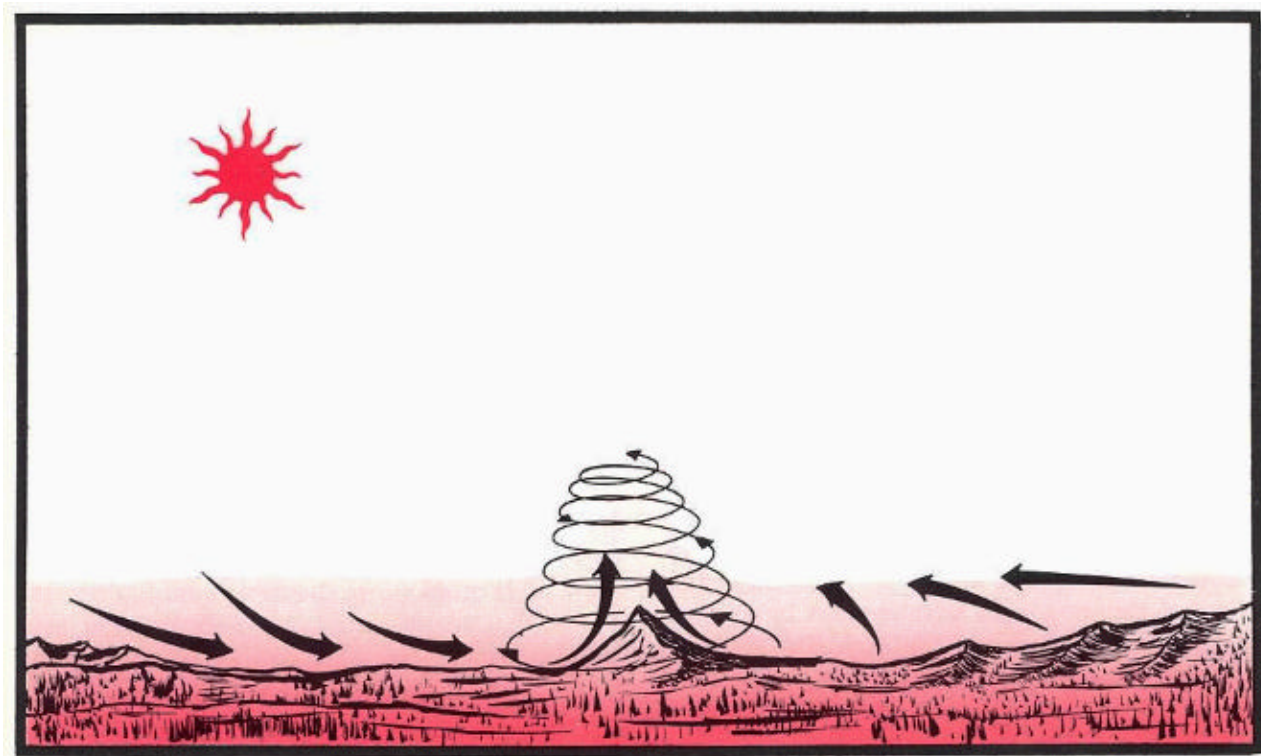
Whirlwinds or dust devils are one of the most common indications of intense local heating. They occur on hot days over dry terrain when skies are clear and general winds are light.

Under intense heating, air near the ground often acquires a lapse rate of 0.2°F. per 10 feet which is about 3½ times the dry-adiabatic rate. The instability is then so extreme that overturning can occur within the layer even in calm air. The superheated air rises in columns or chimneys, establishing strong convective circulations, and drawing in hot air from the surface layer. An upward-spiraling motion usually develops. The spiral is analogous to the whirlpool effect nearly always observed in water draining from a wash basin. The flow becomes spiral because the horizontal flow toward the base is almost invariably off balance.

The lapse rate mentioned in the preceding paragraph is called the autoconvective lapse rate. Greater instability than this may create updrafts spontaneously, but usually a triggering

action initiates the updraft. Updrafts can also begin if the layer acquires only a super-adiabatic lapse rate; that is, a lapse rate less than the autoconvective but greater than the dry-adiabatic rate. However, with superadiabatic lapse rates, quiet surface air actually remains in vertical equilibrium, and becomes buoyant only if it is lifted. In this case, some triggering action must provide the initial impulse upward. One common triggering action is the upward deflection of the surface wind by an obstacle.

It is probable that nearly all updrafts have some whirling motion, but usually this is weak and invisible. The stronger the updraft, the stronger the whirl, because a larger volume of air is drawn into the vortex. The whirling motion intensifies as the air flows toward the center, much the same as the whirling of an ice skater increases as he moves his arms from an extended position to near his body. The whirl becomes visible if the updraft becomes strong enough to pick up sand, dust, or other



Whirlwinds form when sufficient instability develops in a superheated layer near the ground. The latent energy may be released by some triggering mechanism, such as an obstruction or a sharp ridge. Once convection is established, air in the heated layer is drawn into the breakthrough.

debris. The direction of rotation is accidental, depending on the triggering action. It may be either clockwise or counterclockwise.

Whirlwinds may remain stationary or move with the surface wind. If the triggering action is produced by a stationary object, the whirlwind usually remains adjacent to the object. If it does break away, it may die out and another develops over the object. Those whirlwinds that move show a tendency to move toward higher ground. Some whirlwinds last only a few seconds, but many last several minutes and a few have continued for several hours.

The sizes of whirlwinds vary considerably. Diameters range from 10 to over 100 feet, and heights range from 10 feet to 3,000 or 4,000 feet in extreme cases. Wind speeds in the whirlwinds are often more than 20 m.p.h. and in some cases have exceeded 50 m.p.h. Upward currents may be as high as 25 to 30 m.p.h. and can pick up fair-sized debris.

Whirlwinds are common in an area that has just burned over. The blackened ashes and

charred materials are good absorbers of heat from the sun, and hotspots remaining in the fire area may also heat the air. A whirlwind sometimes rejuvenates an apparently dead fire, picks up burning embers, and spreads the fire to new fuels.

Firewhirls

The heat generated by fires produces extreme instability in the lower air and may cause violent firewhirls. Such firewhirls have been known to twist off trees more than 3 feet in diameter. They can pick up large burning embers, carry them aloft, and then spew them out far across the fireline and cause numerous spot fires. At times, the firewhirls move bodily out of the main fire area, but as soon as they do the flame dies out and they become ordinary whirlwinds moving across the landscape.

Firewhirls occur most frequently where heavy concentrations of fuels are burning and a large amount of heat is being generated in a small area. Mechanical forces are often present

which serve as triggering mechanisms to start the whirl. A favored area for firewhirls is the lee side of a ridge where the heated air from the fire is sheltered from the general winds. Mechanical eddies produced as the wind blows across the ridge can serve as the triggering mechanism to initiate the whirl. The wind may add to the instability by bringing in cool air at higher levels over the fire-

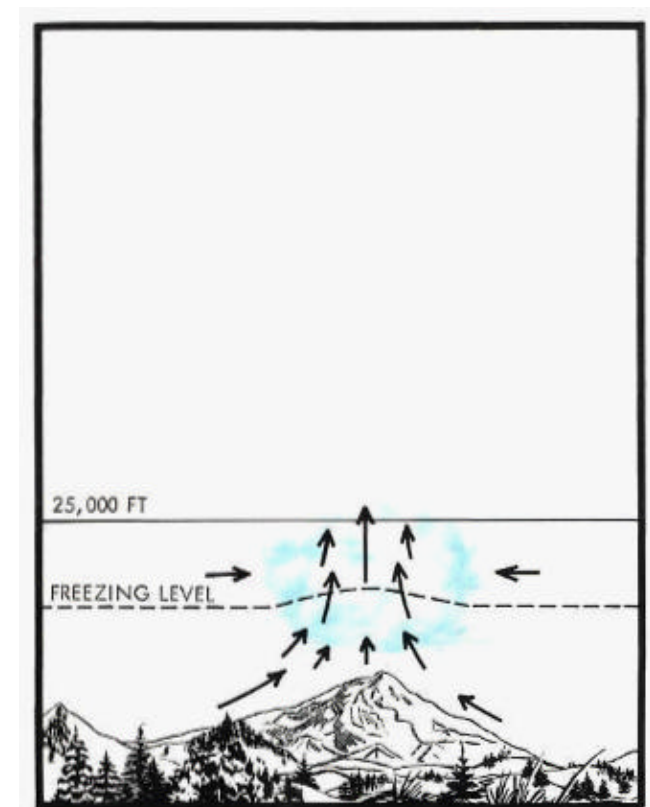
heated air on the lee side. Air streams of unequal speeds or from different directions in adjacent areas can mechanically set off firewhirls in fire-heated air. Firewhirls have also been observed in relatively flat terrain. In these cases the whirls seem to start when a critical level of energy output has been reached by a portion of the fire.

THUNDERSTORM WINDS

Special winds associated with cumulus cloud growth and thunderstorm development are true convective winds. For that reason they will be described here, even though we will consider them again when we look into the stages of thunderstorm development in chapter 10. These winds are (1) the updrafts predominating in and beneath growing cumulus clouds, (2) downdrafts in the later stages of full thunderstorm development, and (3) the cold air outflow which sometimes develops squall characteristics.

There are always strong updrafts within growing cumulus clouds, sometimes 30 m.p.h. or more even if the cumulus does not develop into a thunderstorm. Ordinarily, the air feeding into the cloud base is drawn both from heated air near the surface and from air surrounding the updraft. The indraft to the cloud base may not be felt very far below or away from the cloud cell. A cell that forms over a peak or ridge, however, may actually increase the speed of upslope winds that initiated the cloud formation. A cumulus cloud formed elsewhere that drifts over a peak or ridge also may increase the upslope winds while the cloud grows with renewed vigor. With continued drift, the cloud may draw the ridgetop convection with it for a considerable distance before separating.

If a cumulus cloud develops into a mature thunderstorm, falling rain within and below the cloud drags air with it and initiates a downdraft. Downward-flowing air, which remains saturated by the evaporation of raindrops, is ordinarily warmed at the moist-adiabatic rate. But air being dragged downward in the initial stages of a thunderstorm downdraft is warmed at a lesser rate because of entrainment of surrounding cooler air and the presence of cold raindrops or ice crystals. If this air



The updraft into cumulus clouds that form over peaks and ridges may actually increase the upslope winds that initiated the cloud formation

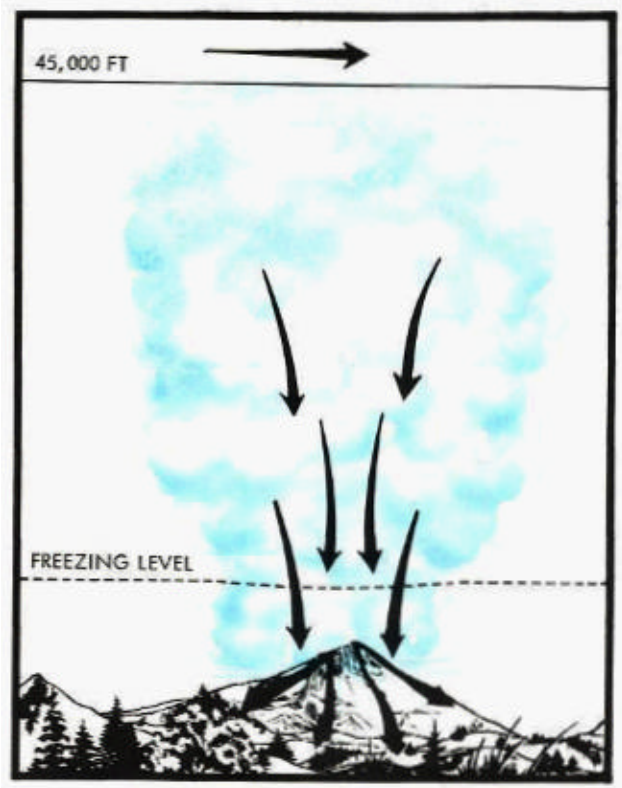
is dragged downward to a point where it is colder than the surrounding air, it may cascade to the ground as a strong downdraft. In level terrain this becomes a surface wind guided by the direction of the general wind and favorable airflow channels. This is known as the first gust and will be treated more fully in chapter 10.

In mountainous terrain the thunderstorm downdraft tends to continue its downward path

into the principal drainage ways. Speeds of 20 or 30 m.p.h. are common, and speeds of 60 to 75 m.p.h. have been measured. If it is dense enough, the air has sufficient momentum to traverse at least short adverse slopes in its downward plunge. The high speeds and surface roughness cause these winds to be extremely gusty. They are stronger when the air mass is hot, as in the late afternoon, than during the night or forenoon. Although they strike suddenly and violently, downdraft winds are of short duration.

Although downdraft winds are a common characteristic of thunderstorms, it is not necessary for developing cumulus clouds to reach the thunderstorm stage for downdrafts to occur. Downdrafts can develop on hot days from towering cumulus clouds producing only high-level precipitation.

Squall winds often precede or accompany thunderstorms in the mountainous West. These storms often cool sizeable masses of air covering an area of a hundred or 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 lower levels, the leading edge—a front—is accompanied by squall winds. These are strong and gusty; they begin and end quickly. They behave much like wind in squall lines ahead of cold fronts, but are on a smaller geographic scale. However, they may travel out many miles beyond the original storm area.



The downdraft in a mature thunderstorm continues out of the base of the cloud to the ground and, being composed of cold air, follows the topography. It strikes suddenly and violently, but lasts only a short time.

SUMMARY

In this chapter on convective winds we have included local winds, which are produced by local temperature differences. Any factors affecting heating and cooling will influence convective winds. These winds will also be affected by the general wind flow. The most familiar convective winds are land and sea breezes, valley and slope winds, whirlwinds, and winds associated with convective cumulus and thunderstorm clouds.

In the land- and sea-breeze system, the local winds are due to land-water temperature differences, which, in turn, produce differences in the temperature of the overlying air. Slope winds are due to temperature differences between slope air

and air over the valley. Valley winds likewise result from temperature differences between valley air and air at the same elevation over the plains. Strong local heating will develop a very unstable layer of air near the surface, and the sudden release of this concentrated energy, usually following a triggering action, may produce whirlwinds.

Thermal updrafts resulting from local heating may produce cumulus clouds, which, under suitable moisture and instability conditions, may develop into thunderstorms. Up-drafts are convective winds characteristic of developing cumulus clouds, but downdrafts are produced in thunderstorms after precipitation

Begins falling from the cloud.

Having considered the general circulation and general convective, winds, we will now turn

to the subject of air masses and fronts, and the weather associated with them.