10

Terrain-Forced Flows

Winds associated with mountainous terrain are generally of two types. Terrain-forced flows are produced when large-scale winds are modified or channeled by the underlying complex terrain. Diurnal mountain winds are produced by temperature contrasts that form within the mountains or between the mountains and the surrounding plains and are therefore also called *thermally driven circulations*. Terrain-forced flows and diurnal mountain winds are nearly always combined to some extent. Both can occur in conjunction with small-scale winds, such as thunderstorm inflows and outflows, or with large-scale winds that are not influenced by the underlying mountainous terrain.

10.1. Three Factors that Affect Terrain-Forced Flows

Terrain forcing can cause an air flow approaching a mountain barrier to be carried over or around the barrier, to be forced through gaps in the barrier, or to be *blocked* by the barrier. Three factors determine the behavior of an approaching flow in response to a mountain barrier:

- the stability of the air approaching the mountains,
- the speed of the air flow approaching the mountains, and
- the topographic characteristics of the underlying terrain.

Unstable or neutrally stable air (section 4.3) is easily carried over a mountain barrier. The behavior of stable air approaching a mountain barrier depends on the degree of stability, the speed of the approaching flow, and the terrain characteristics. The more stable the air, the more resistant it is to lifting and the greater the likelihood that it will flow around, be forced through gaps in the barrier, or be blocked by the barrier. A layer of stable air can split, with air above the *dividing streamline height* flowing over the mountain barrier and air below the dividing streamline height splitting upwind of the mountains, flowing around the barrier (figure 10.1), and reconverging on the leeward side (section 10.3.2). A very stable approaching flow may be blocked on the windward side of the barrier (section 10.5.1).

Moderate to strong cross-barrier winds are necessary to produce terrain-forced flows, which therefore occur most frequently in areas of cyclogenesis (section 5.1) or where low pressure systems (figure 1.3) or jet streams (section 5.2.1.3) are commonly found. Whereas unstable and neutral flows are easily lifted over a mountain barrier, even by moderate winds, strong cross-barrier winds are needed to carry stable air over a mountain barrier. As the speed of the cross-barrier flow increases, the amount of stable air carried over the barrier also increases, with less air channeled around or through the barrier and less air blocked upwind of the barrier.

Strong cross-barrier winds are needed to carry stable air over a mountain barrier.

10.1.1. *Topographic Characteristics of the Underlying Terrain*

A number of terrain characteristics influence both the speed and the direction of terrain-forced flows. The height and length of the mountain barrier can determine whether air goes over or around the barrier. The amount of energy required for air to flow over a high mountain ridge is much greater than that required to flow over a small hill. More energy is required to flow around an extended ridge than around an isolated peak. Thus, high wind speeds are required to carry air over a high mountain range or around an extended ridge. When stable air flows around an isolated peak or the edges of an elongated mountain range, the highest wind speeds are on the hillsides where the flow is parallel to the peak's contour lines (figure 10.2).

The shape of the vertical cross section through the mountain barrier affects wind speed on both the windward and leeward sides. Wind speed increases at the crest of a mountain, with gently inclined triangular-shaped hills producing the greatest increase in speed, flat-topped mesas producing the smallest increases, and rounded mountaintops producing

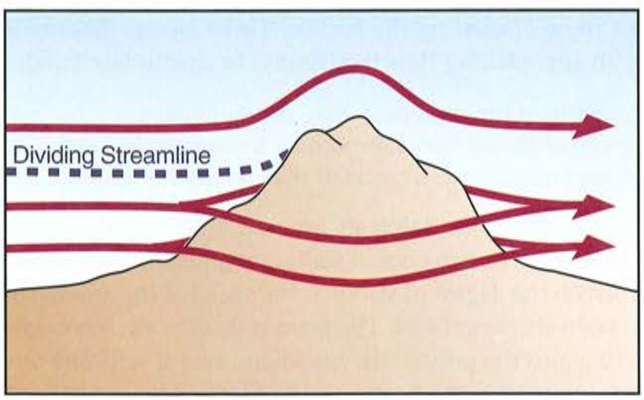


Figure 10.1 The dividing streamline height is the height of the boundary between low-level air, which splits to flow around the barrier, and upper level air, which is carried over the barrier. (Adapted from Etling, 1989)

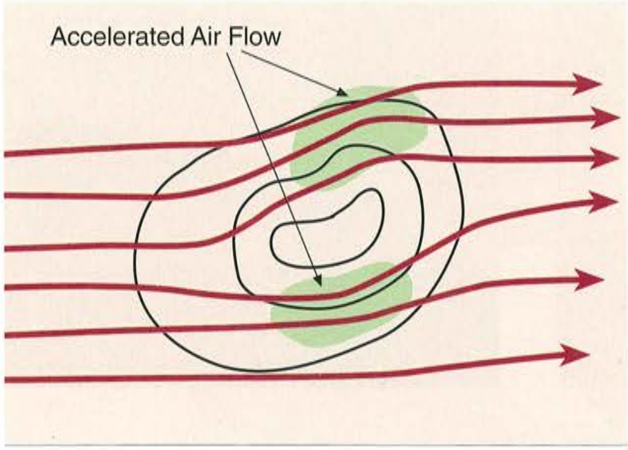


Figure 10.2 Under stable conditions, winds split around an isolated mountain, and strong wind zones are produced on the edges of the mountain that are tangent to the flow. (Adapted from Justus, 1985)

intermediate speedups. Flow separations or *separation eddies* can form over steep slopes or cliffs on either the windward or leeward sides of a barrier. These elongated, horizontal-axis eddies, which can extend along the entire length of the barrier, reduce near-ground wind speeds over the slopes (figure 10.3). The flow above these separation eddies, however, speeds up as it crosses the barrier.

The orientation of a mountain ridge relative to the approaching flow and the curve of the ridgeline (as viewed from above) affect wind direction and speed. Ridgelines that are concave to the windward side and mountain barriers oriented perpendicular to the flow cause flow across the barrier to increase and frequently generate lee waves downwind of the obstacle. Flows approaching barriers with ridgelines that are parallel to, oblique to, or convex to the approaching flow (figure 10.4) change direction to follow the underlying terrain and generate lee waves less frequently. Smaller scale features on a ridgeline can also affect the approaching flow, channeling winds through passes and gaps.

The presence of valleys and basins affects wind speed. Sites low in a valley or basin are often protected from prevailing winds by the confining topography. If winds aloft are strong, however, eddies can form within a valley or basin downwind of a ridgeline, bringing strong, gusty winds to lower elevations.

The roughness of the underlying surface affects wind speed: the rougher the surface, the greater the reduction in wind speed (section 5.2.4). Wind speeds increase when winds move from a rough surface to a smooth surface (for example, from a rough, mountainous area onto a large lake) and decrease when winds move from a smooth surface to a rough surface. The layer in which wind speeds are affected by a roughness boundary deepens with distance downwind from the roughness boundary. An abrupt increase in roughness causes winds to converge, air to rise, and clouds to form. For example, *lake-effect storms* that form in the fall and winter on the eastern shorelines of the Great Lakes result from the abrupt increase in roughness encountered by the prevailing westerly winds. The westerly flow picks up moisture as it moves across the open lakes and then rises when it reaches the shoreline and the rolling hills to the east, producing clouds and locally heavy snowfall.

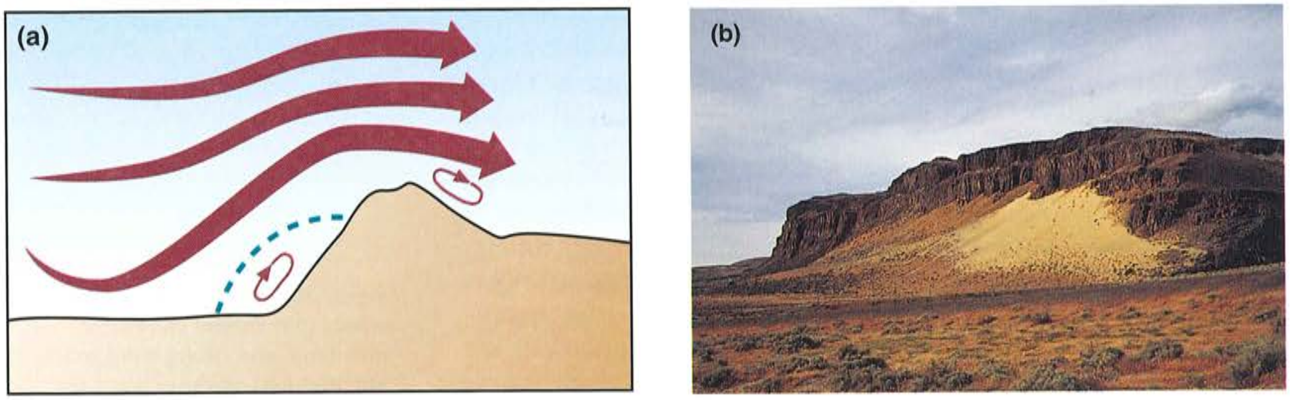


Figure 10.3 Flow separations occur on the windward or leeward faces of steep-sided hills or mountains: (a) A flow separation on the windward side occurs below the dashed green line. A smaller flow separation is also present on the lee side. (b) Sand dunes form where wind speeds drop in separation eddies on the windward face of a hill. This dune formed near Vantage, Washington, on the Columbia River. (Photo © C. Whiteman)

Terrain-forced flows are influenced by the height and length of the mountain barrier, the shape of the vertical cross section through the barrier, the orientation of the barrier relative to the approaching flow, the curve of the ridgeline, the presence of valleys and basins, the roughness of the underlying surface, and terrain obstacles.

Finally, terrain obstacles (mountains, forest patches, shelter belts, individual trees, terrain projections, buildings) generate turbulent *eddies* and/or *wakes* (section 10.1.2) when the approaching flow has sufficient speed. Eddies are swirling currents of air at variance with the main current. A wake forms when eddies are shed off an obstacle and cascade into smaller and smaller scales. Wakes may also form when air splits to flow around rising bubbles or columns of warm air or convective towers in thunderstorms. Wakes are characterized by low wind speeds, but their high turbulence can produce locally gusty winds. They can be seen when air motion tracers (clouds, smoke, or blowing snow or dust) are present and are easily identified when captured by a time-lapse video camera.

Approaching flows that are carried over mountains respond to overall or large-scale features of the topography rather than to small-scale topographic details. In fact, when a low-lying cold air mass is present on the windward and/or leeward sides of the mountain barrier, the approaching flow responds to the combination of the actual terrain and the adjacent air mass. Thus, the *effective topography* that influences the approaching flow can be higher and wider than the actual topography. Terrain-forced flows respond to both landforms (such as valleys, passes, plateaus, ridges, and basins) and roughness elements (such as peaks, terrain projections, trees, and boulders) in complex terrain. Wind speeds can vary significantly between sites exposed to prevailing winds or terrain-forced flows and sites protected from winds by the terrain. Table 10.1 includes information from chapter 6 and provides general rules on where to expect high and low wind speeds.

10.1.2. *Wakes, Eddies, and Vortices*

Eddies can form anywhere in the atmosphere, both at the ground and aloft, in relationship to obstacles and over unobstructed terrain as the result of wind shear or convection. The relative horizontal and vertical dimensions of an eddy are determined by the stability. *Isotropic* eddies (i.e., eddies that have similar vertical and horizontal dimensions) develop in neutral stability, whereas vertically suppressed eddies develop in more stable air, and vertically enhanced eddies form in unstable air.

A wake is an area extending downwind of an obstacle and is characterized by relatively slower wind speeds but increased gustiness. Winds

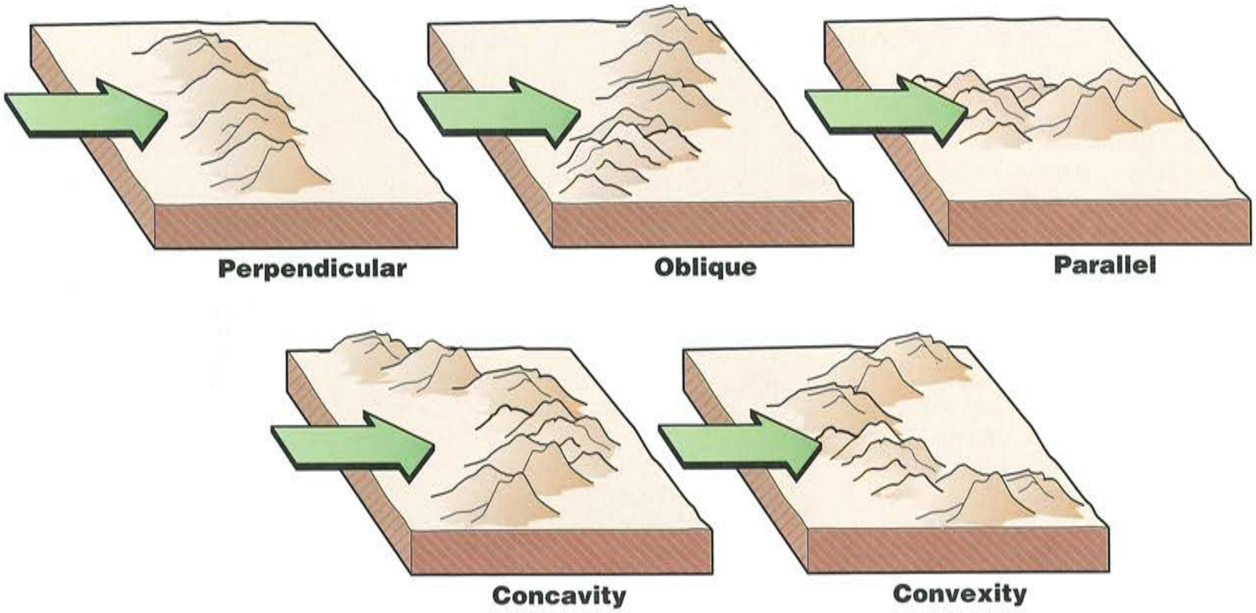


Figure 10.4 The orientation and shape of a ridgeline affect the speed and direction of a flow crossing a mountain barrier. The highest speedups occur over ridgelines that are perpendicular to the flow or have a concavity oriented into the flow. (Adapted from Justus, 1985)

are generally slowed downwind of obstacles to distances roughly 15 times the obstacle height (Angle and Sakiyama, 1991), although velocity reductions in wakes have been detected as much as 60 obstacle-heights downwind.

Vortices, whirling masses of air in the form of a column or spiral, usually rotate around either vertical or horizontal axes. Vertical-axis vortices in the atmosphere range in size from dust devils to tornadoes, waterspouts, hurricanes, and synoptic-scale high and low pressure centers. Horizontal-axis vortices are caused by vertical wind shear and turbulence and are common in the lee of extended terrain obstacles, such as mountain ridges, shelter belts, and snow fences. The reduced wind speed in the lee of shelter belts and snow fences protects crops from damage and highways from drifting snow (figure 5.9). Horizontal-axis vortices can be stretched into vertical-axis vortices by convection when the ground is heated and the atmosphere is unstable.

Eddies are common in mountainous terrain. Their presence is often indicated by the formation of particular clouds or snow cornices. Banner

Table 10.1 Landforms Associated with Strong and Weak Winds at the Ground

Expect high wind speeds at sites:
- located in gaps, passes, and gorges in areas with strong cross-gap pressure gradients
- exposed directly to strong prevailing winds, especially mountain summits, upper windward or leeward mountain slopes, high plains, or elevated plateaus
- located downwind of smooth fetches, as on downwind shores of large lakes or oceans

Expect low wind speeds at sites:
- protected from prevailing winds as at lower elevations in basins or in deep valleys oriented perpendicular to prevailing winds
- located upwind of mountain barriers or in intermountain basins where low-level air masses are blocked by the barrier
- located in areas of high surface roughness, such as forested, hilly terrain

Figure 10.5 The snow on the left of the summit ridge of Handies Peak in Colorado is the remnant of a snow cornice. The general flow, from right to left, causes a horizontal-axis eddy to form on the lee side of the ridge. In this summertime photo, moist air from an afternoon rainstorm is lifted up the slope in the rising part of the eddy where it condenses into a cloud on the left side of the ridge. A hiker on the ridge has cloudy air on one side and clear air on the other. (Photo © C. Whiteman)

clouds form in eddies in the lee of sharp isolated peaks (section 7.1.4.6). Rotor clouds (section 7.1.4.3) form in horizontal-axis eddies associated with trapped lee waves (section 7.1.4.3) some tens of miles downwind of the mountain barrier. Snow cornices can build up in winter when horizontal-axis eddies (figure 10.5) form over the lee slope. Cornices sometimes extend for some distance along the mountain crest (figure 5.10).

Large, generally isotropic vertical-axis eddies can be produced by the flow around mountains or through gaps in the topography as eddies are shed from the vertical edges of the terrain obstruction (figure 10.6). The Schultz eddy on the north side of the Caracena Straits in California's Sacramento Valley and the San Fernando and Elsinore convergence zones on the northeast sides of the Santa Monica and Santa Ana Mountains are examples of large, vertical-axis eddies formed in this way.

10.2. Flow over Mountains

An approaching flow tends to go over a mountain barrier rather than around it if the barrier is long, if the cross-barrier wind component is strong, and if the flow is unstable, near-neutral, or only weakly stable. These conditions are frequently met in the United States because the long, north–south-oriented mountain ranges lie perpendicular to the prevailing westerly winds and the jet stream. Flow over a mountain barrier can be surmised from the presence of certain cloud types, including lenticular clouds, cap clouds, banner clouds, rotors, a foehn wall, a chinook arch, and billow clouds (section 7.1.4). Blowing snow and cornice build-up on the ridge crests in winter and blowing dust at the ridge crests or on the upper lee slope in summer also indicate flow over a mountain barrier. Flow over mountains generates mountain waves and lee waves in the atmosphere and can produce downslope windstorms.

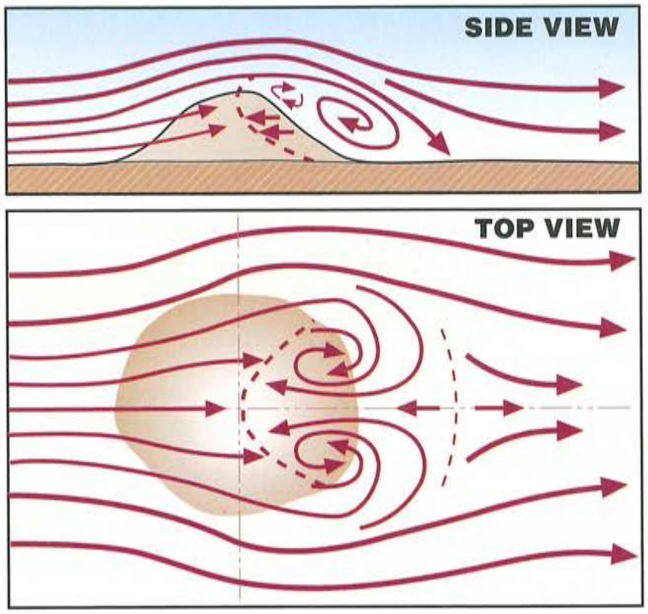


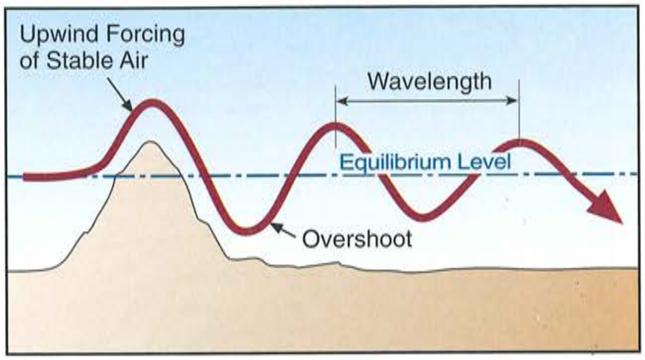
Figure 10.6 Vortex pairs and a wake are generated by the flow around a mountain. (Adapted from Orgill, 1981)

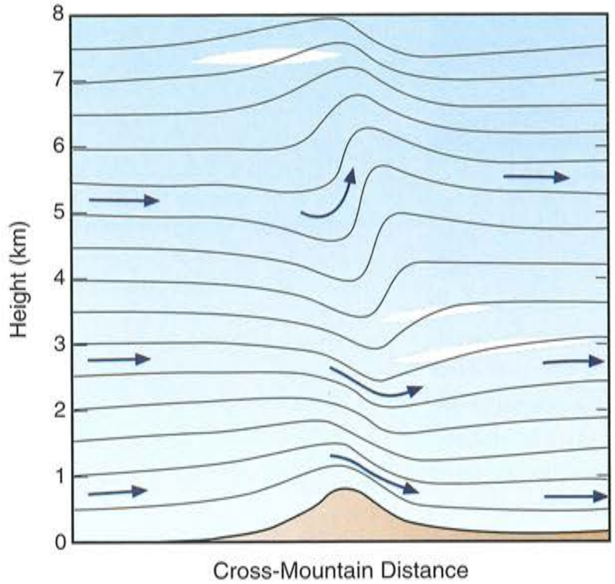
10.2.1. *Mountain Waves, Lee Waves, and Hydraulic Flows*

As stable air flows over a mountain range, *gravity waves* (i.e., vertical undulations or waves in the atmosphere created as gravity acts on local variations in air density) can be generated either over the mountains or in the lee of the mountains. Stable air that is lifted over a mountain barrier cools, becomes denser than the air around it, and, under the influence of gravity, sinks again on the lee side of the barrier to its equilibrium level. The air overshoots and oscillates about its equilibrium height (figure 10.7).

Gravity waves that form over the mountains are called mountain waves. Mountain waves have a tendency to propagate vertically and can thus be found not only at low levels over hills and mountains but throughout the troposphere and even in the stratosphere (figure 10.8). Waves that form in the lee of mountains are called lee waves. Lee waves are often confined or trapped (figure 10.9) in the lee of the barrier by a smooth, horizontal flow above. The two types of waves are collectively called *orographic waves* or simply mountain waves. Classification of waves on a particular day can be difficult because the two types can be present simultaneously and because there is a continuum between the two. In general, mountain waves are found higher in the atmosphere and tend to have longer wavelengths and smaller amplitudes than lee waves.

If there is sufficient moisture in the atmosphere, clouds form in the crests of the waves as air is lifted through the wave in its passage across and downwind of the mountain barrier. Thus, the presence of lenticular clouds is a clear indication of the presence of orographic waves. Larger clouds, such as the chinook arch, are associated with the longer wavelength mountain waves. When no clouds are present, the existence of waves can nonetheless be assumed whenever a stable air mass approaches a significant barrier with a strong cross-barrier wind component.





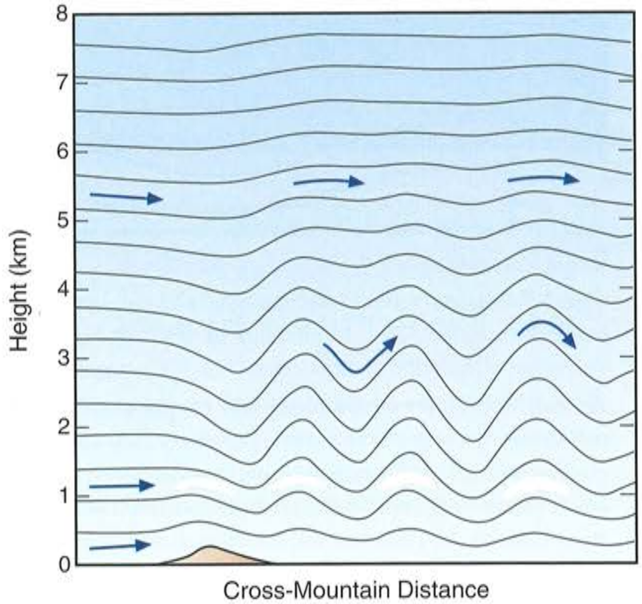


Figure 10.7 (*top*) Stably stratified air that is lifted over mountains oscillates about its equilibrium level on the lee side of the mountain, producing waves. (Adapted from Stull, 1995)

Figure 10.8 (*bottom left*) Vertically propagating mountain waves have their highest amplitudes well above the mountains. High lenticular clouds are often indicators of these waves. (Adapted from Carney et al., 1996)

Figure 10.9 (*bottom right*) Trapped lee waves reach their highest amplitudes in a confined layer on the lee side of the mountains. Regularly spaced low-altitude lenticular clouds are the best indicators of trapped lee waves. (Adapted from Carney et al., 1996)

Thickening of wave clouds and increasing sky coverage are indicative of the increasing moisture that often precedes a frontal passage. The rate of cloud development can be determined from direct observations or from hourly airport observations, to which special remarks on wave clouds (altocumulus standing lenticulars or ACSL) are often appended. The formation of lenticular clouds was discussed in section 7.1.4.4.

Orographic waves form most readily in the lee of steep, high barriers that are perpendicular to the approaching flow. The amplitude of the waves, which decays with distance from the mountain barrier, depends on the initial displacement of the flow above its equilibrium position on the windward side of the mountain. In general, the higher the barrier, the greater the amplitude of the waves. If the flow crosses more than one ridge crest, the waves generated by the first ridge can be amplified (a process called *resonance*) or canceled by the second barrier, depending on its height and distance downwind of the first barrier (figure 10.10).

The basic form of a wave (trapped or vertically propagating) and its wavelength depend on variations in the speed and stability of the approaching flow. When stable air is carried over a mountain barrier, one of three flow patterns will result, depending on the characteristics of the

THE OCTOBER 1997 FOREST BLOWDOWN IN THE CENTRAL ROCKY MOUNTAINS

The largest forest blowdown ever recorded in the Rocky Mountains occurred during a short period on the morning of 25 October 1997. Twenty thousand acres of old-growth trees were blown down on the west side of the Park Range in the Routt National Forest and Mt. Zirkel Wilderness areas northeast of Steamboat Springs, Colorado. Fallen trees were stacked as high as 30 feet (9 m), and hunters in the forest were stranded for as long as two days.

Blowdowns occur most often on the lee sides of mountain barriers, which in Colorado are usually the east side. In this case, however, a strong downslope windstorm developed in easterly flow on the west side of the Continental Divide. The storm was similar to those that occur frequently on the east side of Colorado's Front Range when the amplitude and wind speeds of a deep wave are enhanced by hydraulic flow or lee wave breaking.

The October 25 blowdown in the Park Range was preceded by the development of an intense low pressure storm system in the Four Corners area (where Colorado, New Mexico, Arizona, and Utah meet) on October 24. The counterclockwise winds flowing around the low exposed eastern and northern Colorado to easterly winds that carried moist air up the foothills of the Front Range, producing snowfall up to 4 feet deep with drifts to 15 feet. Wind gusts exceeded 100 mph (45 m/s) for 5 hours at Arapahoe Basin Ski Area (above timberline at 12,500 ft or 3810 m) in the central Colorado Rockies. Wind gusts in the Park Range north of Arapahoe Basin were probably as strong, but according to hunters, were of much shorter duration, lasting only about 30 minutes.

wind profile. If the winds are weak and nearly constant with height, shallow waves form downwind of the barrier. When winds become stronger and show a moderate increase with height, the air overturns on the lee side of the barrier, forming a standing (i.e., nonpropagating) lee eddy with its axis parallel to the ridgeline. When winds become stronger still and show a greater increase with height, deeper waves form and propagate farther downwind of the barrier. Wavelength increases when wind velocities increase or stability decreases.

Orographic waves form most readily in the lee of steep, high barriers that are perpendicular to the approaching flow.

Under certain stability, flow, and topographic conditions, a large-scale instability can cause the entire mountain wave to undergo a sudden transformation to a *hydraulic flow* (figure 10.11). As the name implies, the air flows like river water over a large boulder, with high speeds on the leeward slope, a cavity at the bottom of the slope, and severe turbulence immediately beyond the cavity. A hydraulic flow exposes the lee side of the mountain to sweeping, high-speed, turbulent winds that can cause forest blowdowns and structural damage. Damage is most likely to occur near the base of the lee slope, where the winds are strongest, or slightly downwind of the base of the slope, where turbulence levels are high.

10.2.2. *Downslope Windstorms: The Bora, the Foehn, and the Chinook*

Downslope windstorms occur on the lee side of high-relief mountain barriers when a stable air mass is carried across the mountains by strong

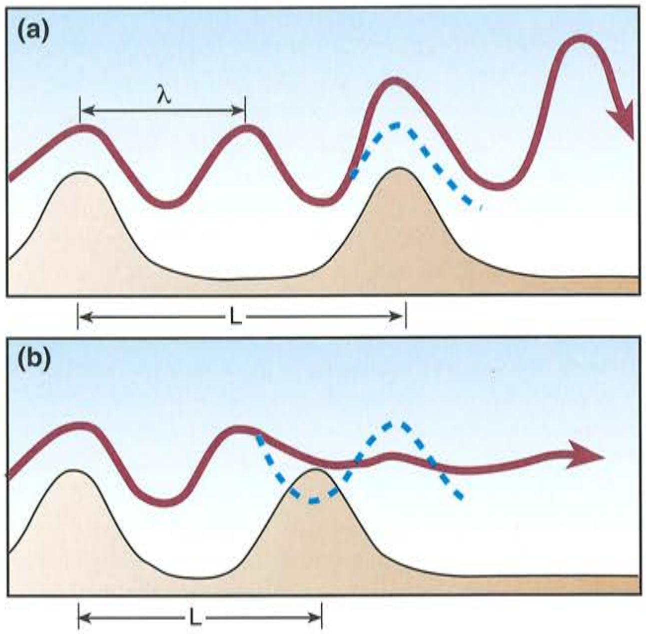


Figure 10.10 Waves can be (a) amplified or (b) canceled by successive ridges, depending on the relationship between the ridge separation distance L and the wavelength λ of the flow. (Adapted from Bérenger and Gerbier, 1956)

cross-barrier winds that increase in strength with height. The strong winds (greater than 50 mph or 22 m/s and sometimes exceeding 100 mph or 45 m/s) are caused by intense surface pressure gradients, with a high pressure center on the upwind side of the barrier and a low pressure trough paralleling the lee foothills. The cross-barrier surface pressure gradient is intensified as the descending air on the lee side of the barrier produces local warming and thus a decrease in pressure at the surface. The pressure gradient may be intensified further if the windstorm coincides with the arrival of a short-wave trough (section 5.1.4), which causes the surface pressure to drop on the lee side of the barrier and to rise on the windward side. The short-wave trough can also cause the winds at mountaintop level to shift direction and become more perpendicular to the barrier. Elevated inversions have been noted in many windstorms where observations were available near mountaintop level. Because elevated inversions are difficult to observe and to forecast, their presence is usually assumed when all other meteorological conditions for windstorms are met. This assumption may result in the overforecasting of windstorms, which is considered preferable to underforecasting.

Downslope windstorms occur primarily in winter on the lee side of high-relief mountain barriers and are associated with large-amplitude lee waves.

Downslope windstorms occur primarily in winter and appear to be associated with large-amplitude lee waves. The descending branch of the first wave reaches the ground at the foot of the slope because the amplitude of the first wave has been increased by resonance (section 10.2.1), by wave trapping (trapping of the vertical energy below a smooth horizontal flow at a given height), or by the development of a hydraulic flow.

Local topography influences the strength of windstorms at a given location. Winds are strong downwind of ridgelines that are high, continuous, and oriented perpendicular to the flow. A ridgeline that is concave on the upwind side also enhances strong winds (section 10.1). Steep leeward slopes, where flow separations form under normal conditions, can cause an acceleration of hydraulic flows.

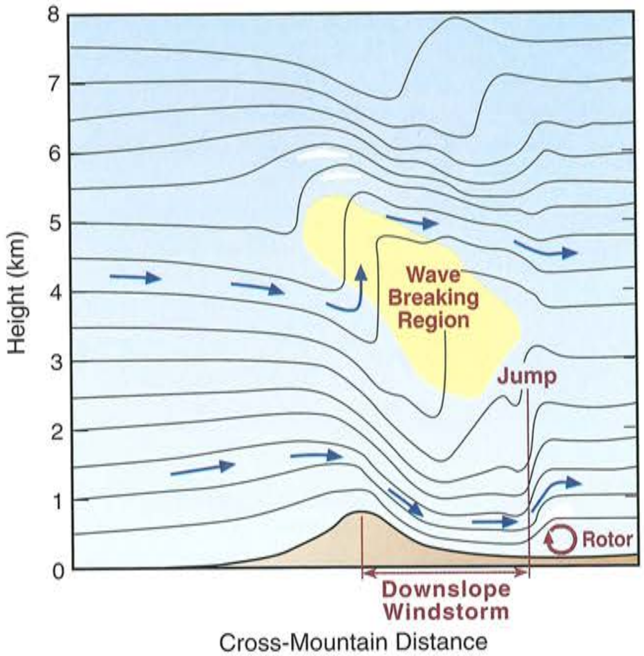


Figure 10.11 Hydraulic flow produces a distinctive flow pattern in the lee of a mountain barrier that is characterized by a region of wave-breaking aloft and a sudden jump in the streamline pattern (*hydraulic jump*) downwind of the barrier. A turbulent rotor cloud may form behind the hydraulic jump. Downslope windstorms may occur during hydraulic flow. (Adapted from Carney et al., 1996)

Downslope winds can bring either cold or warm air into the leeward foothills. A cold downslope wind is called a *bora*, after a wind that brings very cold air to the eastern coast of the Adriatic Sea in Slovenia, Croatia, and Bosnia. The bora originates in an area in central Asia where temperatures are so low that, despite adiabatic warming, the wind is still cold when it reaches the Adriatic coast. The German word *föhn* (or foehn) is used internationally to designate a warm, dry downslope wind. The warming and drying are caused by *adiabatic compression* as air descends the slopes on the leeward side of a mountain range. In the western United States, the foehn is called a chinook, after a Northwest Indian tribe. The term was first applied to a warm southwest wind that was observed at the Hudson Bay trading post at Astoria, Oregon, because it blew from "over Chinook camp" (Burrows, 1901). "Chinook" became more widely used as the adjacent country was settled and is now applied to foehn winds over the entire western part of the United States and Canada (figure 10.12). (Interestingly, the wind at Astoria could not actually have been a foehn wind because the topography southwest of Astoria is not high enough to produce significant adiabatic warming).

In the United States, chinooks are primarily a western phenomena because the topographic relief of the Appalachians is generally insufficient for the production of strong downslope winds. In the Rocky Mountains, chinooks blow most frequently from November through March, bringing relief from long periods of arctic cold east of the mountains (figure 10.13). The gusty, warm winds rapidly melt wintertime snow cover, earning them the name "snoweaters." Four factors contribute to the warmth and dryness of the chinook winds (figure 10.14):

- The air that descends the lee slope is warmed and dried by *compression heating* at the dry adiabatic rate of 5.4°F per 1000 ft or 9.8°C/km

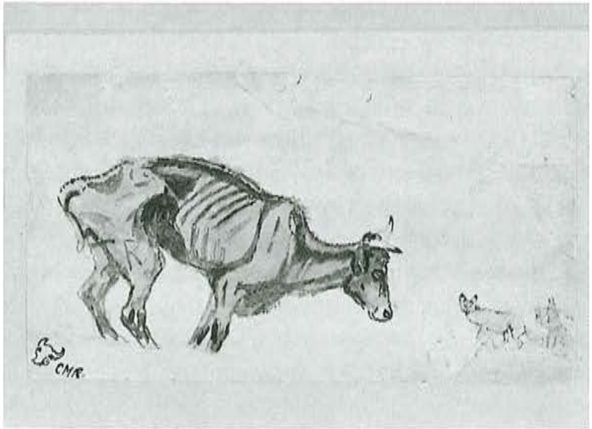


Figure 10.12 (*top*) Foehn winds in the western United States.

Figure 10.13 (*at right*) Waiting for a chinook. (Painting by C. M. Russell, used with permission from the Montana Stockgrowers Association, Helena, Montana) as the air is brought to the lower altitudes and, thus, higher pressures at the base of the lee slope.

- When a deep flow causes air at low levels upwind of the mountain barrier to be lifted up the barrier, latent heating occurs as clouds form and precipitation falls on the windward side, thus warming the air before it descends on the lee side.
- Warm air descending the lee slopes can displace a cold, moist air mass, thus enhancing the temperature increase and the humidity decrease associated with the winds.
- The turbulent foehn flow can prevent nocturnal temperature inversions from forming on the lee side, thereby allowing nighttime temperatures to remain elevated.

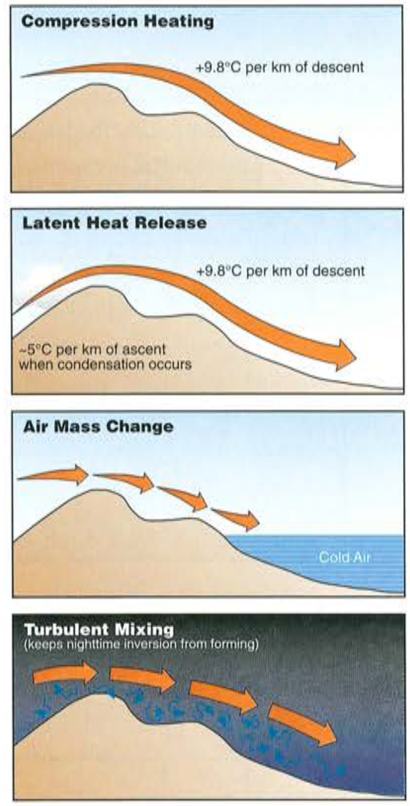


Figure 10.14 Four factors cause the warming and drying associated with chinooks. (Adapted from Beran, 1967)

Downslope windstorms can start and stop suddenly at a given location when changes in the cross-barrier flow component or the stability of the approaching flow cause the wavelength of the orographic waves to change. However, the sudden onset or cessation of winds is usually due to changes in the position of a shallow, cold air layer on the lee side of the mountains that protects the surface from the strong winds. (If the cold air layer is very shallow, strong winds aloft may be heard at the ground). Strong downslope winds can suddenly reach the ground if the upwind edge of the air mass is scoured away by the downslope winds or if waves on the upper surface of the cold air mass depress the surface near its edge. Downslope winds can also reach the ground if prevailing winds aloft drag the cold air away from the mountains or if winds within the cold air mass weaken, allowing it to slide down the topography away from the mountains. The windstorm ends abruptly if the cold air mass sloshes back into place. An abrupt cessation of downslope winds at a given site is called a *foehn pause* or *chinook pause*. Alternating strong wind break-ins and foehn pauses can cause temperatures to oscillate wildly, as illustrated by temperature records for December 1933 in Havre, Montana (see Point of Interest).

Flow speeds in downslope windstorms are highest in a narrow zone along the base of a mountain barrier. For example, cities like Boulder, Colorado, and Livingston, Montana, that are located near the mountain–plain interface of the Rocky Mountains frequently report windstorms. Downslope windstorms also affect less-populated regions along the Rocky Mountain foothills from Canada to southern Colorado but often go unreported because they do not impact large populations. The highest wind speeds occur on elevated mesas or other terrain projections on the edge of the mountains. For example, strong winds are well documented on Table Mesa in southwest Boulder and on Rocky Flats south of Boulder.

Strong, gusty downslope windstorms can cause considerable damage along the mountain–plain interface. Damaging winds rarely extend more than 15 miles (25 km) out onto the adjacent plain, although winds can still be strong at these distances. The strong winds pose a number of hazards. Open fires can be spread rapidly by the winds, and local flooding can result from the rapid melting of snow cover. Smoke, windblown dust, and strong gusts can cause poor driving conditions. In built-up areas, damage to roofs and fences, and windows broken by windblown objects are common. Buildings under construction and mobile homes are particularly susceptible to wind damage. Aviation hazards include turbulent

Flow speeds in downslope windstorms are highest in a narrow zone along the base of a mountain barrier, where the strong, gusty winds can cause considerable damage.

rotors that develop below the crest of lee waves and in the lee of the hydraulic flow cavity. Damage caused by downslope windstorms may be exacerbated by vertical-axis eddies, sometimes called *mountainadoes*. Mountainadoes develop when vertical wind shear near the ground produces horizontal eddies that are then stretched vertically to rotate about a vertical axis. If these vertical eddies are advected with the mean wind, they produce strong horizontal shears and wind gusts that are much more damaging than the general prevailing winds.

Because most of North America is within the zone of prevailing westerlies, chinook winds usually occur on the eastern sides of U.S. mountain ranges. They can, however, occur on the western sides of mountain barriers when upper level winds are from the east. The *Santa Ana winds* in the Los Angeles Basin and the *Wasatch winds* in Utah are familiar examples of northeasterly or easterly chinook winds.

The Santa Ana winds develop in the late fall and winter when a high pressure center in the Great Basin reverses the normal pressure gradient that causes winds to blow off the ocean in Southern California. The high

BATTLE OF THE CHINOOK WIND AT HAVRE, MONTANA

The thermograph record in figure I shows a series of extreme temperature variations that occurred at Havre, Montana, during the week of 15–22 December 1933 when chinook winds repeatedly broke through a shallow cold air mass. A southwest chinook struck Havre on 16 December, causing temperatures to rise 27°F (15°C) in only 5 minutes as wind velocities increased from 5 to 30 mph (2 to 13 m/s). The chinook continued until the morning of 18 December, when temperatures fell 40°F (22°C) as cold arctic air spread over Havre from the north and east. Chinook break-ins occurred again for several hours on 19 December and during the nights of 20–21 December and 21–22 December, with sudden temperature variations of 20–30°F (11 to 17°C).

The weather maps in figure II illustrate the changes in frontal position that produced alternating periods of cold and warm air temperatures at Havre. In figure IIa, a developing low pressure storm system in Alberta was accompanied by warm and cold fronts. Warm chinook winds descended the lee slopes of the Rockies in Saskatchewan, Montana, and Wyoming to push cold air eastward away from the Rockies and immerse Havre in the warm air behind (west of) the warm front. Meanwhile, a surge of cold air was moving southward along the eastern slopes of the Rockies to the west of the low. In figure IIb, the cold air surged as far south as Colorado and immersed Havre in cold air. The cold air dome was dammed from westward movement by the Rockies, as indicated by the stationary front.

The conceptualized weather maps in figure III illustrate a typical break-in of warm chinook winds on the east slope of the Rockies. In figure IIIa, a stationary front indicates the western edge of a cold air layer that is dammed against the eastern slope of the Rockies. In figure IIIc, a decrease in the easterly component of winds within the cold air mass allows it to slide southeastward down the slope of the plains, exposing the western edge of the plains to westerly chinook winds.

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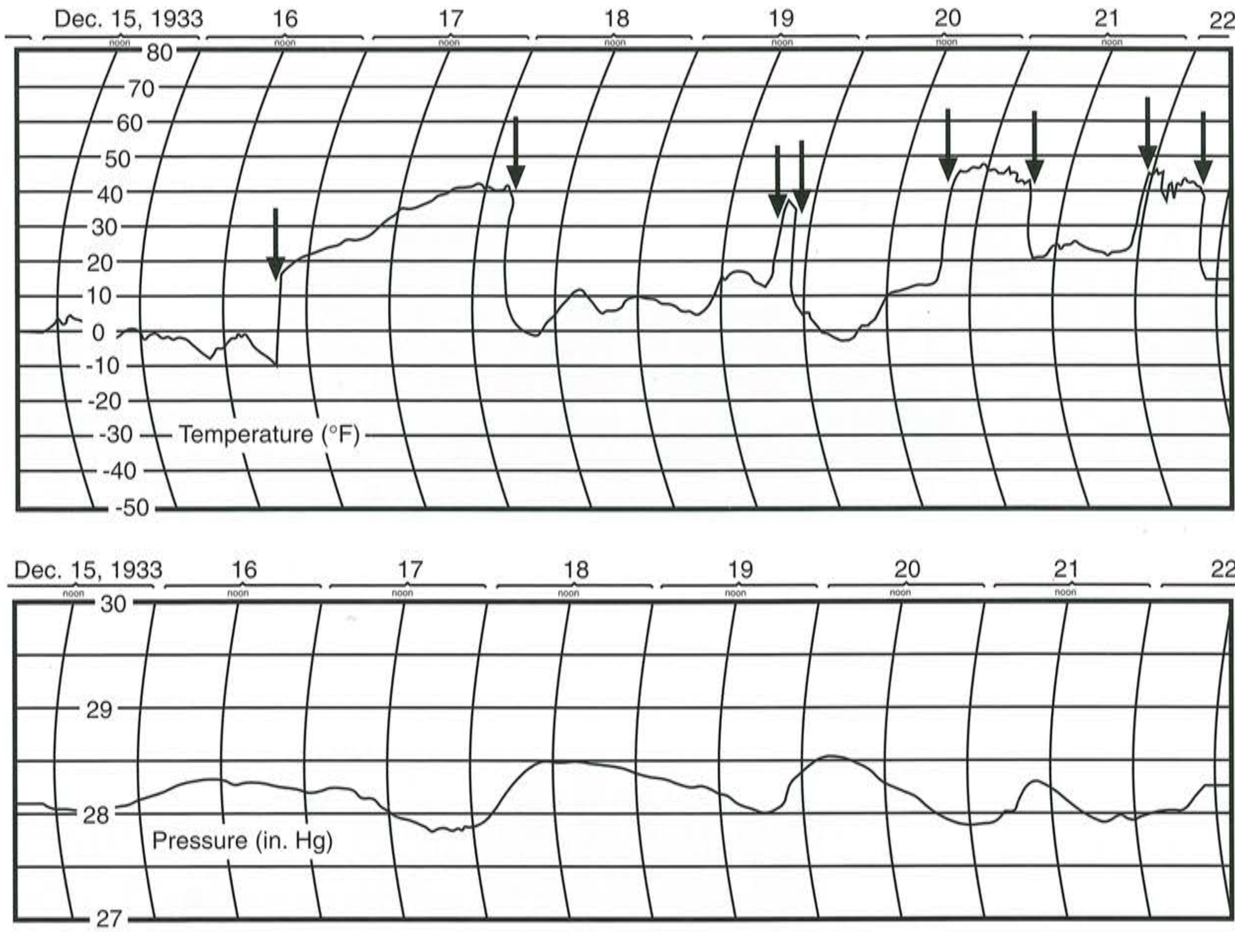

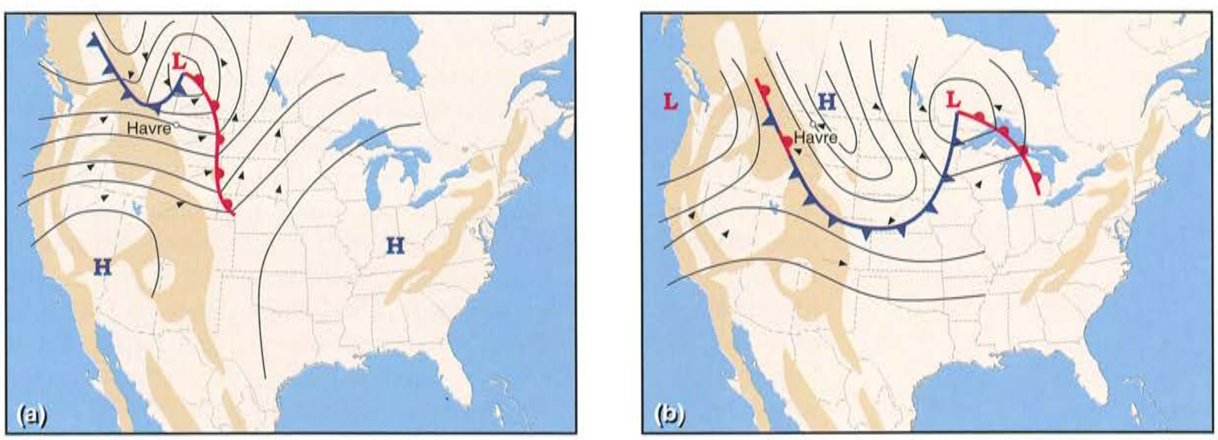

Figure I (*top*) A series of eight abrupt temperature changes occurred as warm chinook winds and shallow cold air masses alternated at Havre, Montana, during the week of 15–22 December 1933. Surface pressures increase when the cold air is in place. (Adapted from Math, 1934)

Figure II (*bottom*) Surface weather charts for (a) 18 December 1933 at 0100 UTC and (b) 19 December 1933 at 0100 UTC show the battle of the chinook wind at Havre, Montana. (Adapted from Glenn, 1961)

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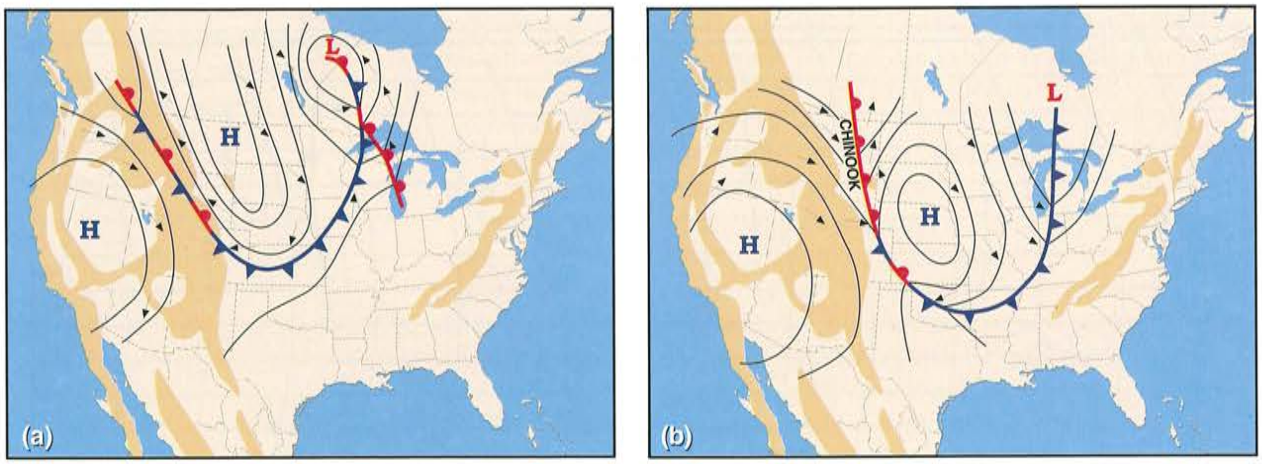


Figure III Conceptualized surface weather charts illustrating the break-in of chinook winds as an arctic air mass is carried eastward away from the mountains. (Adapted from Glenn, 1961)

pressure center channels air south and west out of the Great Basin around the southern end of the Sierra Nevada, over the Mojave Desert, and through low passes (e.g., Cajon Pass or Bannon Pass) into the Los Angeles Basin (figure 10.15). The synoptic pressure gradients leading to the Santa Ana winds (figure 10.16) are distinctive and easily recognized by forecasters. These foehn winds can be strong and gusty and can last for several days. Because the winds are dry, they desiccate the vegetation and can whip *wildfires* into firestorms that threaten property and lives.

Wasatch or *canyon winds* occur on the west side of Utah's Wasatch Mountains between Ogden and Provo when a strong east–west pressure gradient develops across the north–south Wasatch Range east of Salt Lake City. The easterly winds blow into a low pressure center in the Great Basin. Canyon winds appear to be of two types. The more common are winds that come through gaps in the ridgeline of the Wasatch Range and are channeled down the major canyons, producing localized strong and gusty winds at the canyon mouths. Less common are windstorms that affect a more or less contiguous zone along the foothills, with the strongest winds usually occurring in the lee of the major unbroken north–south sections of the Wasatch ridgeline. This second type of wind is produced by strong mountain waves or hydraulic jumps or combinations of the two. An example of a synoptic weather chart for a Wasatch wind episode is shown in figure 10.17. The wind strengths have been correlated with 850-mb height differences between Fort Bridger, Wyoming, and Salt Lake City. The greater the height difference (Salt Lake City has the lower height), the stronger the winds. (Similar "East winds" are found on the west side of the Cascades in Washington (figure 10.12).).

Gap winds that develop along the Pacific coast in Washington, Oregon, British Columbia, and southeast Alaska are sometimes enhanced by foehnlike flows when a strong east–west pressure gradient exists at and above mountaintop level between the coast and the inland areas, with low pressure along the coast. Winds are channeled westward through gaps in the coastal ranges and blow out over the Pacific Ocean or over the waterways of the Inside Passage. They are especially strong downwind of low passes or where major river valleys issue onto the seaways, such as below

WINDSTORMS AT BOULDER, COLORADO

- Boulder has between 2 and 16 windstorms per year.
- Windstorms are most frequent in Boulder from November through March (see figure I).
- Windstorms are somewhat more likely to occur at night than during the day.
- The typical windstorm lasts about 8 hours.
- The 11 January 1972 windstorm caused over 2.5 million dollars of damage in Boulder, with much of the damage in mobile home parks, housing construction areas, and residential areas where strong winds and/or blowing materials caused glass breakage and roofing damage.
- Historical records of windstorm damage in the Boulder area extend back to the 1860s.
- Parked railroad cars were reportedly blown off the tracks between

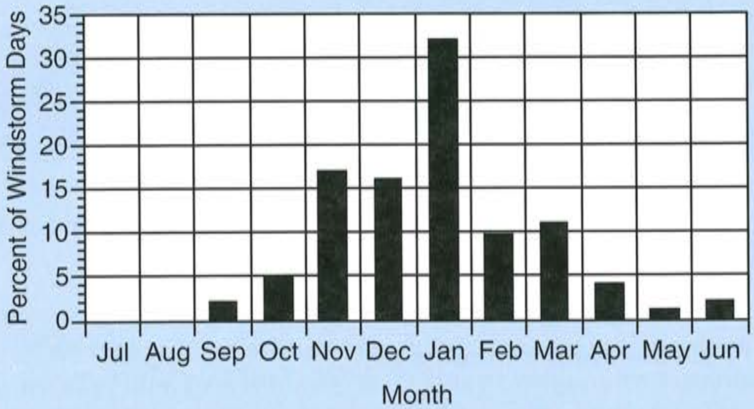
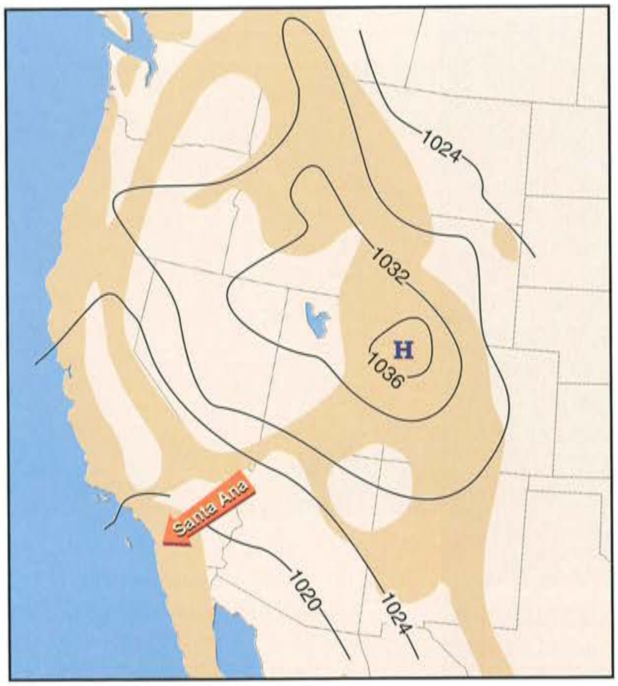


Figure I Damaging downslope windstorms at Boulder, Colorado, occur most often in January, but have relatively high frequencies of occurrence in the months November through March. (Adapted from Whiteman and Whiteman, 1974)

Figure 10.15 Santa Ana winds blow out of the Mojave Desert through the Santa Clara River Valley and through Cajon and Banning Passes toward the Pacific Ocean. Wildfires are very difficult to control when these strong, gusty, dry winds are at their strongest. (Adapted from Rosenthal, 1972)



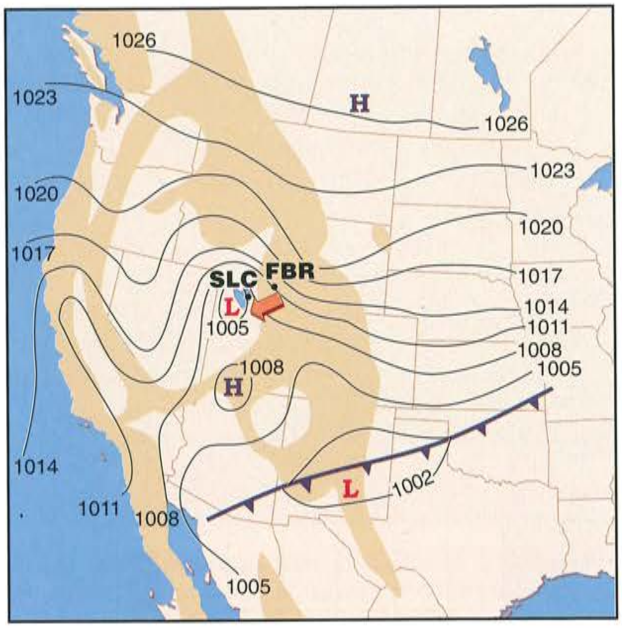


Figure 10.16 (*left*) The Santa Ana winds occur when a strong pressure gradient develops between the Great Basin and the coast of Southern California. (Adapted from Ahrens, 1994)

Figure 10.17 (*right*) This weather chart for 16 May 1952 at 1200 UTC shows the synoptic situation for a Wasatch wind event in which winds reached 95 mph (43 m/s) north of Salt Lake City. Wind strength is proportional to the pressure gradient between Salt Lake City (SLC) and Fort Bridger (FBR). Isobars are sea level pressures in millibars. (Adapted from Williams, 1952)

the Columbia River Gorge, in the Strait of Juan de Fuca, and in and below the Fraser River and Taku River Valleys.

10.3. Flow around Mountains

A flow approaching a mountain barrier tends to go around rather than over the mountains if

- the ridge line is convex on the windward side,
- the mountains are high,
- the barrier is a single isolated peak or a short range,
- the cross-barrier wind component is weak,
- the flow is very stable, or
- the approaching low-level air mass is very shallow.

The Rockies and the Appalachians are so long that flow around them is uncommon. Flow around subranges or isolated peaks occurs most frequently. Examples are found in the Aleutians, the Alaska Range, the Uinta Mountains of Utah, the Olympic Mountains in Washington, and around isolated volcanoes in California, Oregon, Washington, and Alaska.

Flow around mountains can sometimes be detected by watching clouds. On the windward side of the mountain barrier, low clouds are often seen extending, like a scarf, from one edge of the mountain barrier to the other. The clouds may pour over the barrier at its periphery and cascade down the lee slopes. Clouds and precipitation may also form in the lee of the barrier as the air currents that split around the barrier reconverge.

THE ALPINE FOEHN AND THE WESTERN CHINOOK

In the Alps, the characteristics of the foehn and the towns and valleys frequently affected by it are well known. The residents of the Alps are attuned to the foehn, which is a topic of daily discussion and interest. The foehn is valued for the warmth and the partly cloudy skies that it brings, a welcome respite from the cold and foggy conditions that persist for much of the winter on the north side of the Alps generally and on the Swiss Plateau in particular. However, the foehn is also associated with changes in human behavior, health, or psychological outlook and is an active area of research in Alpine medicine. Reports of headaches, low spirits, general ill-being, or depression are common, with some residents reporting that symptoms develop before the onset of strong winds. Various medications are available to treat the general feeling of malaise, described as feeling "foehnig." Schools may close during the foehn, and the effects of the winds can be used as a medical defense in court cases.

In the American West, chinooks are associated with strong, warm winds, although they are also responsible for many warm winter days when winds are not particularly strong and gusty. Some residents complain of headaches during windstorms, but other symptoms are rarely linked to the storms.

10.3.1. *Barrier Jets*

Barrier jets form when a stably stratified, low-level flow approaches a mountain barrier of fairly limited length with an edge or low pass on its left side. The flow, unable to go over the barrier, turns to blow parallel to the longitudinal axis of the barrier. Barrier jets always turn to the left in the Northern Hemisphere as they approach the barrier (figure 10.18). Thus, if a low-level stably stratified air mass approaches a north–south barrier from the west, it is turned to the left and becomes a south wind; a flow approaching from the east is also turned to the left and becomes a north wind.

Barrier jets turn to the left in the Northern Hemisphere as they approach a mountain barrier.

Barrier jets are observed infrequently in North America because of the extreme length of the major mountain barriers, but they have been reported on the west side of the Cascades and Sierra Nevada (figure 10.19), on the east sides of the Rockies and Appalachians, and on the north side of the Brooks Range. Although few observations are available, barrier jets probably develop around the smaller subranges of the Rockies. For example, there is anecdotal evidence that a southerly barrier jet sometimes occurs on the west side of the Colorado Rockies when the prevailing flow is from the west, carrying air around the north (i.e., left) side of the main crest and over the lower lying terrain of southern Wyoming.

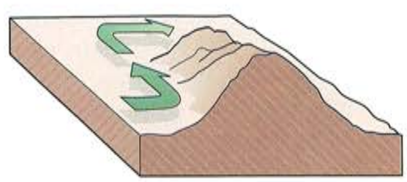


Figure 10.18 A barrier jet occurs when winds are turned to the left to flow around the edge of a mountain barrier.

10.3.2. *Flow Splitting and Convergence Zones*

Flow splitting occurs when the length of a mountain barrier is limited and both the right and left sides are open, as is the case with isolated peaks and roughly circular mountain ranges such as the Olympics or the Alps.

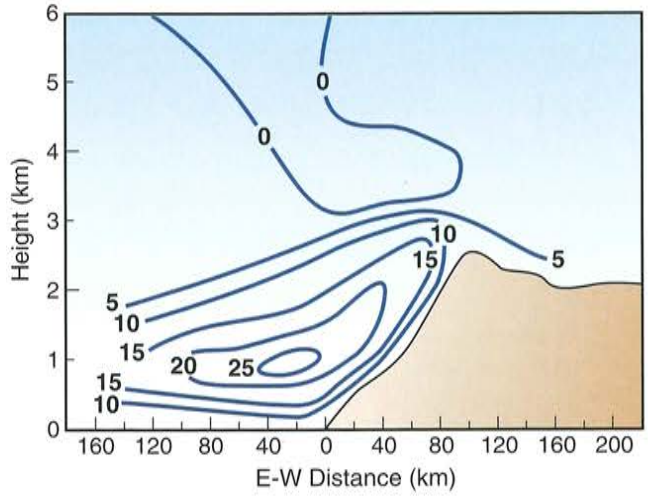


Figure 10.19 East–west cross section through a barrier jet that formed on the west side of the Sierra Nevada. Isolines of constant wind speed (*isotachs*) are labeled in m/s and show the high wind speeds at the core of the flow. The southerly flow from the Sierra Nevada barrier jet is thought to be responsible for unusually high precipitation on the east side of the Trinity Alps at the north end of the Sacramento Valley. (Adapted from Parish, 1982)

Part of the flow is carried around the left side of the barrier, part around the right side (figure 10.20). The flow around the left side is usually somewhat enhanced by the barrier jet.

The two branches of the split flow converge on the barrier's lee side, causing rising motions above the convergence zone and thus fog, cloudiness, and precipitation. The downwind location of the convergence zone shifts in response to changes in the direction of the approach flow and may also vary diurnally when conditions are influenced by diurnal mountain winds.

In North America, a well-known convergence zone is located in the lee of the Olympic Mountains of western Washington. Onshore flows split around the near-circular Olympic Mountains, flowing north through the Strait of Juan de Fuca and south through the low-lying Willapa Hills. The flows converge on the east side of the Olympics in the Puget Sound area, where low-level moisture is often present. Rising motions above the zone of low-level convergence produce clouds and precipitation. It is difficult to predict the exact location of the Puget Sound convergence zone because it shifts as the onshore wind direction changes, but it often affects operations at SeaTac International Airport south of Seattle. Convergence zones also exist in the lee of the Santa Monica and Santa Ana Mountains of Southern California.

A flow that splits around a barrier converges in the lee of the barrier. Fog, cloudiness, and precipitation may form in the convergence zone.

10.3.3. *Frontal Blockages and Postfrontal Accelerations*

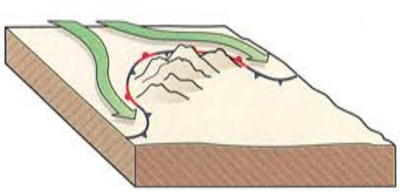


Figure 10.20 A conceptualized view of a cold air mass that approaches the Alps from the northwest and splits to flow around the eastern and western edges of the mountain ridge.

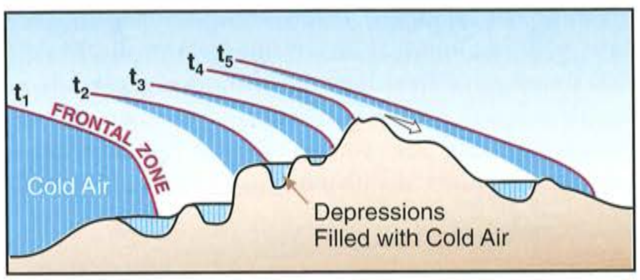
The forward movement of an approaching cold front can temporarily slow as the cold front is forced to ascend a mountain range. As the depth of the cold air mass behind the cold front becomes deeper and deeper, the cold air may begin to invade the valleys, or the front and the cold air behind it may split and flow around the edges of the barrier. If the amount of cold air flowing around the edges of the barrier is small, or if the depth or rate of accumulation of cold air is sufficient, the front will continue to ascend the barrier until it eventually reaches and exceeds the mountain ridge height. The front will then suddenly accelerate as the cold air flows over the mountain ridge and pours down the lee side of the mountain, moving the front past the barrier and exposing the lee slopes to sudden cold air outbreaks (figure 10.21).

A cold front slows as it ascends a mountain barrier and accelerates when the cold air behind it flows over the barrier.

Channeling and distortion of fronts and postfrontal cold air outbreaks frequently occur in the Alps because of the limited horizontal extent and distinct edges of the range (figure 10.22). Similar wind systems are features of the wind *climatology* around the Pyrenees and subranges of the Alps (figure 10.23). Although flow splitting and postfrontal cold air incursions are not common around the long Rocky Mountain and Appalachian ranges, they can occur there when a cold air mass approaches an isolated subrange. A dense network of surface weather stations and radiosonde launch sites in the Alps provides ample documentation of these flows, whereas the low density of weather stations in the United States does not generally provide sufficient data.

10.4. Flows through Gaps, Channels, and Passes

Strong winds are often present in *gaps* (major erosional openings through mountain ranges), in channels between mountain subranges, and in mountain passes. The winds are usually produced by *pressure-driven channeling*, that is, they are caused by strong horizontal pressure gradients across the gap, channel, or pass. The pressure gradient may be im-



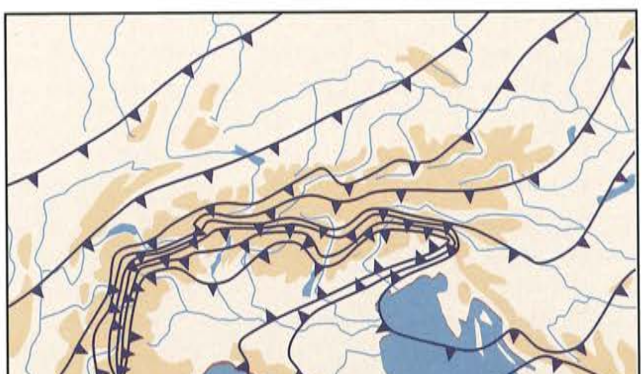


Figure 10.21 (*top*) Fronts are retarded as they move up mountain barriers but accelerate once the barrier is surmounted. The symbols t_1, t_2, t_3, t_4, and t_5) indicate regular time intervals and illustrate the slowing of the front as it climbs the barrier.

Figure 10.22 (*bottom*) Successive 3-hour frontal positions as a cold front approaches and passes the Alps, from 4 March 1982 at 0000 UTC to 5 March 1982 at 1800 UTC. The front slows as it encounters the Alps (dark tan shading). Cold air intrudes into the major Alpine valleys and surges around the eastern and western edges of the Alps. In this case, the stronger surge is around the east side of the Alps. (Adapted from Steinacker, 1987)

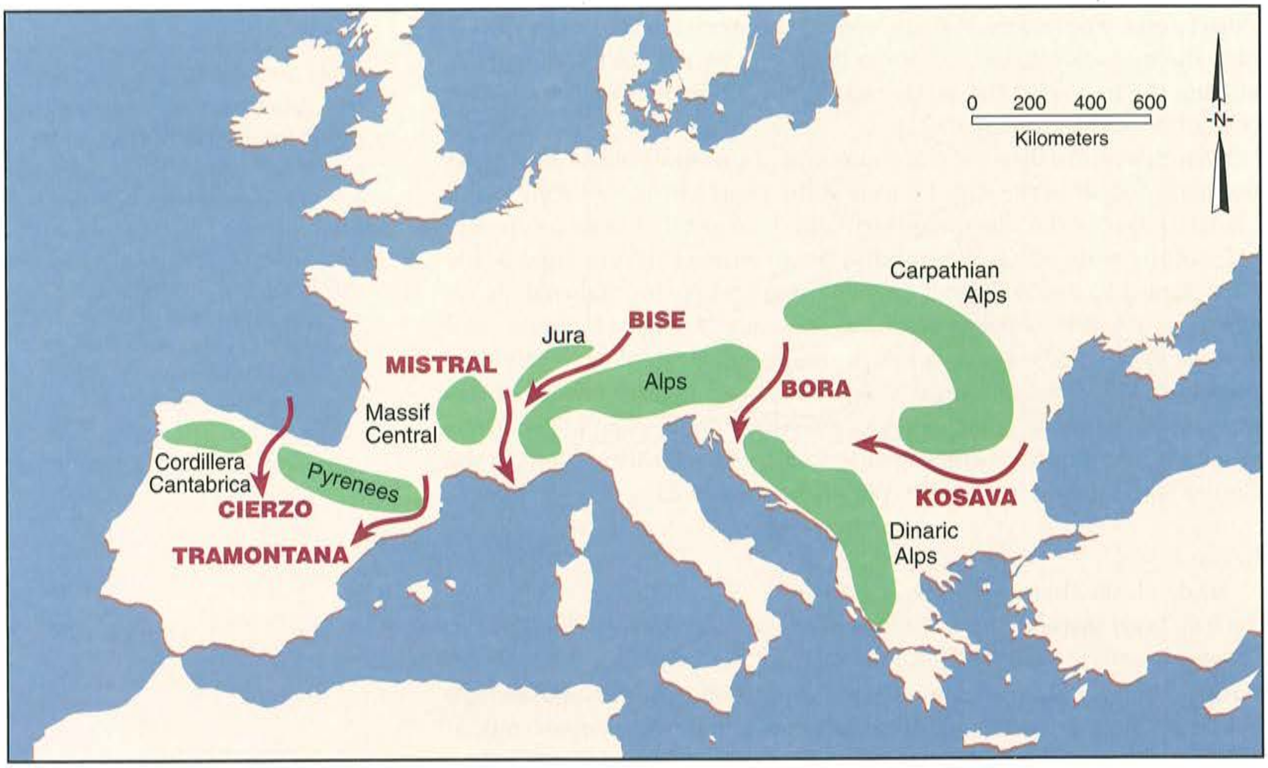


Figure 10.23 Flows between the major mountain barriers in Europe have been given special names. The Bora, flowing between the Alps and the Dinaric Alps, is also a downslope wind descending to the Adriatic coast. (Adapted from Wanner and Furger, 1990)

posed on the terrain by traveling synoptic-scale pressure systems or may result from differences in temperature and density between the air masses on either side of the opening. The differences are usually caused by regional-scale processes, but may also result from smaller scale processes, such as cold thunderstorm outflows. The strongest gap winds occur when synoptic-scale pressure gradients are superimposed on regionally developed pressure gradients. Because the physical processes are similar for gaps, channels, and passes, the phenomenon will be illustrated primarily for gaps.

10.4.1. *Flows through Coastal Mountain Ranges*

Strong winds in a gap, channel, or pass are usually pressure-driven, that is, they are caused by a strong pressure gradient across the gap, channel, or pass.

Regional pressure gradients occur frequently across coastal mountain ranges because of the differing characteristics of marine and continental air masses. In winter, the coldest and densest air is found over the elevated interior of the continent, whereas temperatures on the ocean side are moderated by the open ocean. When the pool of cold air in the interior becomes deep enough, it spills over the lowest altitude gaps, producing strong winds that descend toward the coast (figure 10.24). Further deepening of the cold pool increases the flow through the gap and allows air to flow over low ridges and eventually through higher altitude passes. When low clouds are present, the cloud mass fills the gap or pass and descends like a waterfall into the adjacent valley (figure 10.25). Because major channels or gaps are generally lower in altitude and wider than passes, they carry the strongest flows and largest air volumes. In summer, continental interiors are warmer than the adjacent ocean, and pressure gradients develop, forcing air through the gap from the ocean to the interior. Thus, it is typical for regionally driven pressure gradients in gaps to reverse between winter and summer, with the direction of the flow determined by the direction of the pressure gradient, usually toward the ocean during winter and toward the continental interior during summer.

Perhaps the best known gap wind in the United States develops in the Columbia River Gorge, the main east–west gap through the Cascade Range that connects the Pacific to the interior Columbia Basin (figure 10.26). The pressure difference between stations on the west and east ends of the gorge determines the direction and strength of the winds. In summertime, the Pacific High is off the coast of Oregon and Washington and a low is located over the Columbia Basin, producing an eastward flow that brings marine air up the gorge. The flow is moderately strong and blows opposite to the direction of the river current, producing the dependable up-gorge winds and river waves that are well known to windsurfers. In wintertime, when cold air collects in the Columbia Basin, the pressure gradient is reversed and strong, cold, easterly flows, called Columbia Gorge winds, come down the gorge. The easterly flows undercut warm, precipitating air masses that are being lifted up the western slope of the Cascades, enhancing precipitation and setting the stage for severe ice storms in the lower gorge. Summer and winter winds similar to the gap winds in the Columbia River Gorge also occur through low passes in the Cascade Mountains, including the Snoqualmie, Naches, and Stampede Passes.

Gap winds blow through other major gaps along the West Coast of North America, including the Caracena Strait near San Francisco, the Strait of Juan de Fuca near Seattle, the Fraser Valley near Vancouver, the Stikine Valley near Wrangell, the Taku Straits just south of Juneau, the Copper River Valley just east of Cordova, and the Turnagain Arm south of Anchorage. As in the case in the Columbia River Gorge, the summer

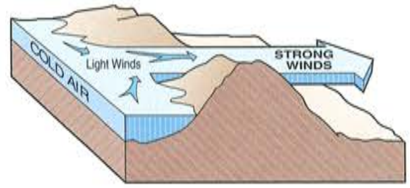


Figure 10.24 Flows can accelerate as they come through mountain passes.

Flows through coastal ranges usually reverse seasonally, flowing toward the ocean during winter and toward the continental interior during summer.

Figure 10.25 A waterfall cloud over Mackinnon Pass (1154 m), a major pass through the Southern Alps of New Zealand on the Milford Track. Clouds are pouring westward across this low pass from the Clinton River Valley into the Arthur River Valley. (Photo © C. Whiteman)

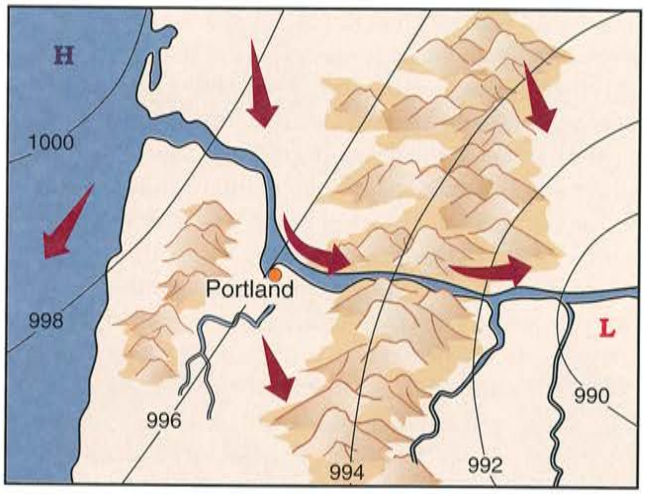


Figure 10.26 Winds are channeled through the Colombia River Gorge. The consistent up-valley flow produces well-developed waves on the Columbia River and wind velocities that are ideal for wind surfing.

inflows through these gaps are relatively weak compared to the strong wintertime outflows.

Wintertime gap winds that blow out onto the Pacific Ocean or onto the waterways of the Inside (or Inland) Passage pose a serious hazard to shipping, especially to slow-moving barges pulled by tugboats. (The Inside Passage is nonetheless the preferred shipping channel because of strong winds and high waves on the open Pacific Ocean.) Of particular concern to barge traffic is the passage between Cross Sound at the north end of Chichagof Island and the anchorage at Cordova, where barge traffic moves onto the open ocean but is also subject to gap winds at the outlet of the Copper River. Winds along the Inside Passage are complicated by channeled winds that frequently blow along the complex of marine channels, canals, or straits, driven by pressure gradients that develop along the length of the channels.

Forecasts of gap winds are usually based on a forecast of pressure differences across the gap. Forecasters monitor surface pressure differences across the gap from surface weather stations and watch the development of synoptic pressure systems. Forecasts of wintertime gap winds along the coast of western Canada and southeast Alaska, for example, rely on the pressure measurements in the cold air mass on the east side of the Coast Range in the Rocky Mountain Trench (e.g., at Whitehorse) and on pressure changes along the coast as low pressure centers or troughs approach the coast from the west. When a low approaches from the west, winds north of the low have an easterly component that strengthens the gap wind flow.

10.4.2. *Flow into Heat Lows*

Confined desert or semiarid basins or plateaus have little soil moisture and vegetation and therefore experience little evaporation and large sensible heat fluxes during daytime. A large sensible heat flux results in a deep, warm convective boundary layer above the surface and relatively low pressures in the basin or on the plateau. A low pressure formed in this way is termed a *heat low*. Horizontal pressure gradients between the heat low and its higher pressure surroundings drive winds that bring air from the surroundings into the heat low, especially at low altitudes where the horizontal pressure gradients are strongest. A heat low contributes to the summertime low pressures over the Columbia Plateau that bring marine air up the Columbia River Gorge. The best known and most strongly developed heat low in the United States forms in summer over the Mojave Desert and the south end of the Great Basin. Winds blow into this low pressure area from the Gulf of California to the south and across passes in the Sierra Nevada to the west and north. Networks of wind machines have been installed on the Tehachapi and Altamont Passes to harness these winds for the production of electrical power.

10.4.3. *The Venturi (or Bernoulli) Effect*

When a valley or other channel has a substantial pressure gradient along its length and a topographic constriction at some point along the channel, air is accelerated through the constriction by the pressure drop across the constriction (figure 10.27). Acceleration through a terrain constriction is called the *Venturi* or *Bernoulli effect*. The flow speed can be roughly estimated by assuming that the mass of the flow is conserved through the channel, so that speed increases when the cross section of the flow narrows and decreases when the cross section widens. When the pressure gradient across the constriction is weak or the constriction provides a near-total blockage, air may pool behind the constriction rather than accelerating through it (section 11.4.2.1).

10.4.4. *Forced Channeling*

Most winds through gaps, channels, and passes are driven by the difference in pressure from one side of the gap to the other. These winds blow across the pressure contours from the area of high pressure to the area of low pressure. Some winds, however, result when strong flows aloft under neutral or unstable conditions are channeled by landform features, such as parallel mountain ranges or valleys. *Forced channeling*, in contrast to pressure-driven channeling, requires the downward transfer of momentum into the channel from winds aloft that are blowing parallel to pressure contours (figure 10.28). Forced channeling often occurs with foehn windstorms and at high altitudes in the mountains, especially at mountain passes. Forced channeling is described in more detail in section 11.3.2.

Forced channeling is the transfer of momentum from winds aloft into a terrain channel.

10.5. Blocking, Cold Air Damming, and Obstruction of Air Masses

A mountain barrier can not only channel an approaching flow over, around, or through gaps and passes in the barrier, it can also block an approaching flow or it can obstruct a shallow, stable air mass that develops on one side of the barrier, channeling it along the foot of the barrier and preventing it from crossing the barrier. Blocking, cold air damming,

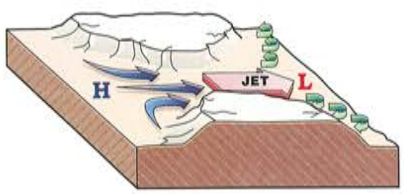


Figure 10.27 The Venturi effect causes a jet to form as winds pass through a terrain constriction and strengthen.

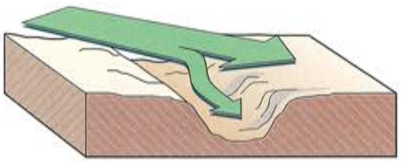


Figure 10.28 Forced channeling occurs when upper winds are brought down into valleys from aloft and turned to flow along the valley's longitudinal axis.

Blocking, cold air damming, and obstruction all affect stable air masses and occur most frequently in winter.

and obstruction all affect stable air masses and occur most frequently in winter.

10.5.1. *Blocking*

A low-level, stably stratified flow approaching a mountain barrier can be stopped or blocked at the barrier as dense air builds up over the windward slope of the mountains, causing the formation of a local high pressure (figure 10.29). The resulting mesoscale pressure gradient can counteract the synoptic-scale pressure gradient that drives the approaching flow, sometimes blocking or stalling the approaching air mass. Blocking occurs frequently in winter when cold, stable air masses are common and may persist for extended periods. Smoke and pollution emitted into a blocked layer can be trapped for days in the stalled flow.

The blocked flow upwind of the barrier is usually shallower than the barrier depth. Air above the blocked flow layer may have no difficulty surmounting the barrier and may respond to the effective topography (section 10.1.1), that is, the combination of the orographic barrier and the blocked cold air mass on its upwind side.

The horizontal extent of a blockage upwind of a barrier depends on the height of the barrier, the stability of the approaching flow, and the latitude. The higher the barrier and the more stable the flow, the greater the horizontal extent. As latitude increases, the horizontal extent of the blockage decreases. If an isothermal air mass approaches a 3250-ft-high (1000-m) barrier at 40°N latitude, blocking extends about 125 mi or 200 km upwind of the barrier. The blockage upwind of a mountain half as high (1625 ft. or 500 m) extends about half this distance. The speed of the approaching flow does not affect the horizontal extent of a blockage.

In basins on the windward side of mountain ranges, a blocked air mass can form a more or less horizontally stratified pool of cold air across the entire basin. Blockages occur frequently in winter in the Intermountain Basin west of the Rocky Mountains, with individual episodes persisting for periods as long as 10 days. The strength of the pool (i.e., the temperature difference between the bottom and top of the pool), as well as its depth, vary as cold air drains into the basin from the surrounding mountains and as cold or warm air is advected into the basin by synoptic flows. The strength of the pool increases when warm air is advected above the pool and decreases when cold air is advected above the pool.

The onset and cessation of blocking can be abrupt, occurring over periods as short as one hour and causing sudden changes in dispersion conditions for smoke or air pollution discharges. In winter, blocking is influenced primarily by synoptic conditions, with the pressure gradient, and therefore blocking, increasing when a synoptic ridge passes the barrier and decreasing when a synoptic trough (often a short-wave trough) passes the barrier. If the upwind stable layer is shallow or only weakly stable, blocking ends when daytime convection upwind of the mountain is sufficient to break through the stable layer. The flow in the resulting neutral layer has no difficulty surmounting the mountain barrier. The blockage may reform at night when longwave radiation loss cools the atmosphere.

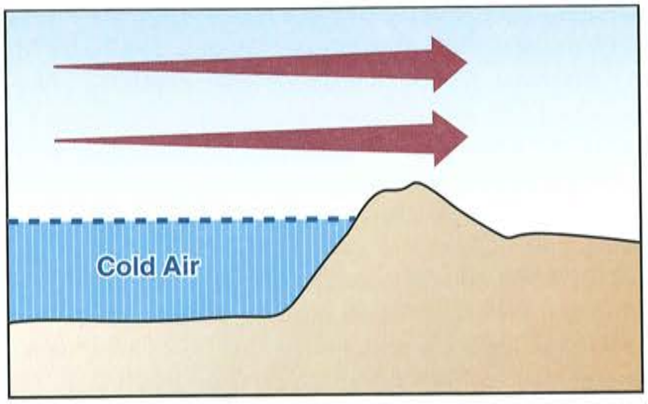


Figure 10.29 Flow blocking produces a layer of stagnant air at low levels upwind of the mountain barrier.

10.5.2. *Cold Air Damming*

Cold air damming is a wintertime phenomenon that occasionally affects the east sides of the Rocky Mountains and Appalachians, resulting in persistent low temperatures, precipitation, and ice storms. A cold, shallow, high pressure center surges southward on the east side of the mountain barrier, a trough or low approaches the mountains from the west, and a pressure gradient is established that initiates an easterly flow over the east side of the barrier. Pressure-driven winds lift the cold air partway up the east slope of the mountains. The air, cooled even further by adiabatic ascent, cannot rise all the way up the barrier and becomes trapped over the east slope. The shallow, wedge-shaped air mass effectively alters the terrain configuration of the barrier to flows that approach it from the east. Warm, moist air to the east of the mountains, also carried westward by the pressure gradient, is less stable than the cold air and can be lifted up and over the cold air dam, causing stratiform or cumuliform cloudiness and precipitation many miles upwind of the barrier. The warm precipitation can turn to snow, freezing rain, or sleet as it falls through the entrenched cold air mass below. The cold air dam becomes more stable when rain or snow falls from above and evaporates into the cold dry air within the dam. The type and location of precipitation along the foothills can be difficult to predict when a cold air dam is present because the position and depth of the cold air mass relative to the mountain barrier can vary with time.

10.5.3. *Obstruction of Air Masses*

In contrast to blocking and cold air damming, which result from the dynamic lifting and cooling of an approaching flow, *obstruction* of an air mass is a static process that can occur on either the upwind or downwind side of a barrier. An air mass develops or moves along one side of a barrier and is unable to surmount the barrier, either because the air mass is too shallow or because the cross-barrier wind component in the air mass is too weak (figure 10.30). Obstruction occurs most frequently along long barriers that prevent cold, stable, very shallow, or dense air masses from flowing around the barrier.

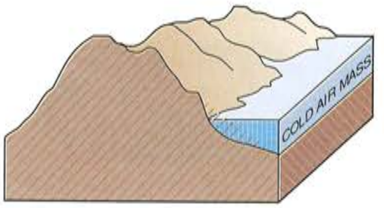


Figure 10.30 Mountain chains make an effective barrier for shallow air masses. The cold air mass here represents a shallow polar air mass on the east side of the Rocky Mountains.

When shallow air masses are common near a long mountain range, the range can be a climatic divide. In North America, the Rocky Mountains and the Appalachians are both climatic divides, confining polar and arctic air masses that form in the higher latitudes and travel south across central Canada to the plains between the two mountain ranges. Because temperatures in the air masses are low and the density is high, the air masses form high pressure centers. Air to the south of the high flows toward the west, forcing the shallow cold air up the Great Plains to the eastern foothills of the Rocky Mountains. Lifting is generally insufficient, however, to carry the shallow dense air over the Rockies. Air to the north of the high flows eastward but is obstructed by the Appalachians. Because the flow on the leading edge of the high pressure area is easterly, the cold shallow air mass hugs the eastern foothills of the Rockies and the Sierra Madre Oriental as it moves south into the Gulf of Mexico and, in some cases, as far south as the Bay of Campeche and the Gulf of Tehuantepec in Central America.

10.6. On the High Plains: The Low-Level Jet

A strong southerly or southwesterly low-level wind, called the low-level jet (LLJ), frequently occurs during nighttime over the Great Plains, which slopes downward from the foothills of the Rocky Mountains to the Mississippi River. This nocturnal flow has a distinctive *jet wind speed profile* (figure 10.31) with maximum speeds often in excess of 35 mph (16 m/s) only 800–1500 ft (250–450 m) above the ground. The LLJ is focused in a narrow band with typical widths of 200–400 miles and extends from the Gulf of Mexico northward into the central Great Plains. The LLJ typically begins at sundown and persists through the night into midmorn-

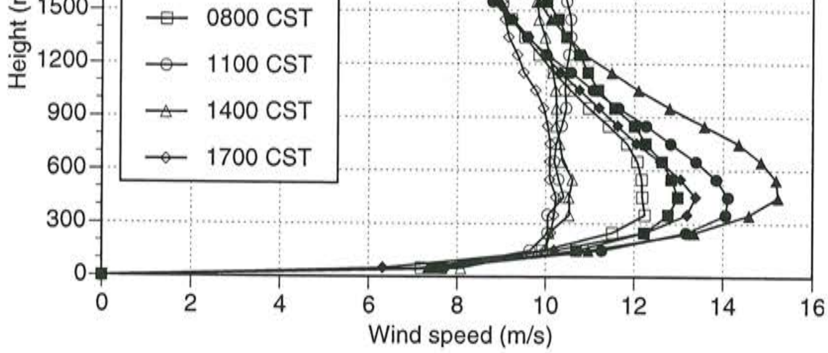


Figure 10.31 Wind speed profiles vary diurnally over the southern Great Plains when the low-level jet develops. The average jet reaches its maximum intensity of 15–20 m/s (34–45 mph) at an elevation of 400–500 m (1300–1600 ft) above the ground at 0200 CST. (From Whiteman et al., 1997)

THE LLJ AND BIRD MIGRATION

Birds and insects use the southerly LLJ to migrate northward in the spring; they often wait for nights when the LLJ is particularly strong or seek flight altitudes that provide the best tailwinds. Many species of birds migrate at night, a fact driven home to meteorologists when a larger network of vertically pointing meteorological radars was installed in the midwest in the early 1990s. This network reported anomalous nighttime winds that were ultimately traced to radar returns from migrating birds. Because birds are better radar scatterers than the atmosphere, the radars measure the ground speed (i.e., the sum of the bird's flight speed, about 10 m/s, and the speed of the jet) of the birds rather than the speed of the air current. In the spring, when the birds fly north with an LLJ tailwind, the radar-reported wind speeds are about 10 m/s too high. In the fall, when the birds fly south into an LLJ head wind, they choose altitudes where the LLJ is relatively weak. Radar-reported wind speeds are about 10 m/s too slow during this time.

ing. Although it can occur in any season, the LLJ is most frequent in the summer half of the year.

A southerly or southwesterly low-level jet occurs frequently during nighttime over the Great Plains.

Although the mechanisms that cause the LLJ to form are still being debated, the correspondence between the spatial extent of the phenomenon and the spatial extent of the Great Plains suggests that topography plays an important role. One theory is that the southeasterly flow around the Bermuda–Azores High over the southern Great Plains is channeled by the slope of the Great Plains and the Rocky Mountains to produce a southerly flow. Another theory suggests that an inclined temperature inversion over the sloping plains provides a thermodynamic driving force. A final factor, independent of topography, that is widely acknowledged as a contributor to the formation of the LLJ is the decoupling of the boundary layer winds from the frictional drag of the earth's surface once surface-based inversions begin to form in the early evening.

The low-level jet can be easily distinguished from the barrier jet (section 10.3.1) that forms east of the Rockies. The LLJ blows from the south and extends far to the east of the Rockies, whereas the barrier jet blows from the north and is closer to the mountain barrier.