4

WINDS

Chapter Highlights

- ✓ Learn about the all-powerful jet stream
- ✓ Understand how wind interacts with mountainous terrain.
- ✓ Become an expert on mountain, valley and glacier winds.

Strong winds are probably the most inconvenient and potentially life threatening meteorological phenomena that you will encounter in your travels among the mountains. Strong winds can occur under a very broad range of conditions; basically in combination with any other type of weather scenario. In addition, by virtue of the mountain environment, high winds can generate extremely poor visibility by lifting dirt or snow from off of the ground, or blowing snow horizontally during a storm. As was noted in the previous chapter, strong winds are also responsible for the 'apparent' reduction in air temperature that mountain travelers frequently experience (i.e.-wind chill). Being able to recognize high wind situations and avoid them if possible is a skill that you should attempt to master.

Jet Stream Winds

As a mountain traveler you should care about the jet stream for two simple reasons: first, wind speeds frequently exceed 70 m/s (150 mph), in short, they are the strongest persistent winds found in the lower atmosphere. Secondly, almost all big storms develop, mature, and dissipate in the vicinity of the jet stream.

We have been using the singular so far, however there are actually three jet streams: the arctic, polar (Figure 4.1), and sub-tropical . The good news is that these 'brutes' are for the most part confined to elevations above 5,000 m (16,400 ft or 500 mb). Furthermore the current suite of computer weather models that are in use, forecast the position and strength of the jet streams quite well. Now for the bad news: you do not want to be at a high elevation when one of these jet streams is in the neighborhood.

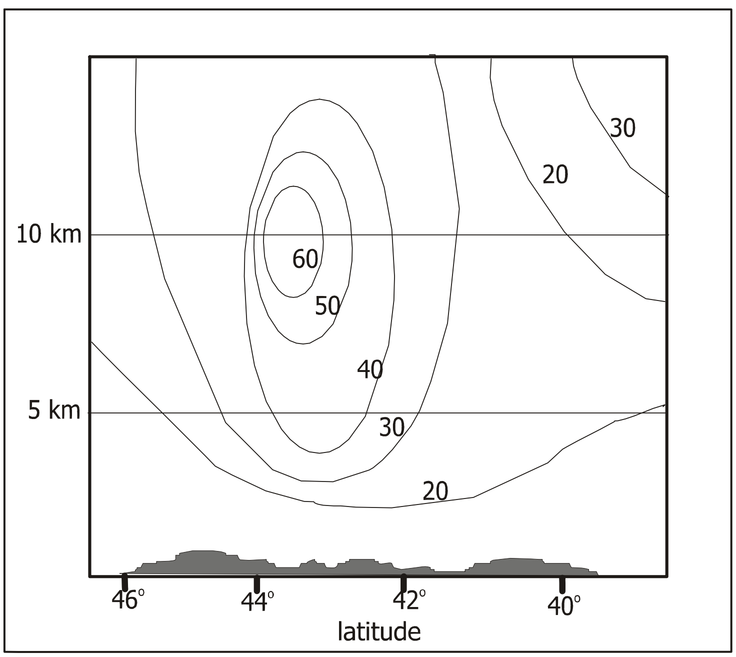


Figure 4.1- Idealization of the polar jet. Contours represent wind speeds in ms^{-1}, with air moving into the page.

The arctic jet (we often drop 'stream') is the least understood and the most recent one to be discovered. It is generally located between 60°-80° N, and a elevation of 500-400 mb This is not a permanent feature, in other words there are days when the arctic jet is present, but many days when it is absent. In fact, it is absent more days then it is present. Likewise, it is not continuous around the globe, it may appear in the Canadian Arctic but not over Scandinavia, or vice versa. Wind speeds in the arctic jet stream typically range from 30-45 m/s (60-90 mph).

The polar jet is the one that most people are familiar with in large part because it has a very large influence on mid-latitude weather. However, the polar jet is not fixed in any one position, it makes a sinuous path across the Northern Hemisphere (there is also one in the Southern Hemisphere), but is most frequently found between 35°-60° N. In addition, it is weaker in the summer months when the temperature difference between high-latitudes and mid-latitudes weakens. There are two regions of the globe where winds in the polar jet reach maximum speed; off the east coast of Asia, near Japan, and off the east coast of North America, near Newfoundland. During the winter both of these regions have very large temperature contrasts due to the proximity of very cold land masses to the west and relatively warm oceans to the east. The level of strongest winds within the polar jet stream occur between 8-10 km (300-200 mb). Wind speeds range from 40-90 m/s (90-200 mph), and on occasions even higher.

The sub-tropical jet develops between the mid-latitudes and the tropics. It generally lies between 20°-30° N and is an important factor in the generation of thunderstorms over the southern USA in the summer and it also produces strong winds over the Himalaya during the cooler months of the year. The sub-tropical jet forms in the descending portion of the Hadley cell, therefore the level of maximum winds tends to be centered around 10-11 km (~200 mb). Wind speeds are similar to those of the polar jet, but can at times exceed 110 m/s (240 mph) <u>over a limited area.</u>

One of the more difficult concepts to grasp is the fact is that all three of these jet streams vary in strength and position from week-to-week or day-to-day. In fact if you were to examine daily weather charts of upper level winds, you would find that on some days only one of the jets is discernable while on another day two or even three of the jets are evident. By way of example, on any given day the polar jet may be strong over the western North Pacific, while the arctic jet is absent and the sub-tropical jet is weak over the North Pacific but strong over the Gulf of Mexico. Two days later the polar jet may have weakened considerably while the sub-tropical jet has moved further north and intensified. As you can imagine, there are many different possible combinations.

The horizontal dimensions of these jets vary greatly, they are typically 500-800 km (300-500 mi) wide and vary in length from a thousand kilometers to several thousand kilometers. Embedded within the jet stream are moving zones of maximum winds known as *jet streaks*. I<u>n a statistical sense, the higher you climb the more likely you are to encounter strong winds</u>. When the sub-tropical jet lies over the Himalaya in the winter, strong winds are the rule not the exception. However, if strong winds are present at 7,000 m (23,00 ft) for example, it does not necessitate strong winds at 4,000 m (13,100 ft) There are a number of factors that determine whether jet stream winds extend to lower elevations.

<u>Discussion</u>: Suppose you are on a three day climb of Mt. Rainier, and your' camped at 3400 m (11,150 ft) on the Tahoma Glacier. You poke your head out of the tent at dawn on summit day, and you observe upper level cirrus clouds moving rapidly towards the northeast-your clue that upper level winds are strong (polar jet). But what about wind conditions on the Summit, 1000 m (3,280 ft) above? Your best indicator is to look for blowing snow higher up the mountain, if you do not observe any,

then the winds are probably light, at least at present.

Keep in mind that upper level clouds in the preceding scenario act as tracers. Blowing snow, dust, or fast moving clouds simply help us monitor the wind at a distance. Strong winds can develop without any cloud development of course, but when clouds are present use them to estimate mid-level wind speeds. When you are in the mountains it can be difficult determining whether or not a jet is in the neighborhood. One indicator is the presence of long rows or streaks of cirrus clouds. In this case the jet is parallel to the long-axis of the clouds. It is also very helpful if you know a prior what the major storm patterns are for the area you are in. For example, in the Cascades major storms move in from the northwest-to-southwest, it would be rare to have persistent strong winds from the southeast. The Mountain Weather Survey chapters of this book are intended to provide you with this type of information.

Interaction of Wind with Terrain

Blocking

Mountains of any size impede or disrupt the movement of air, the larger the mountain or mountain range, the larger it's impact on the flow of air over, through, or around them. One very common and important class of terrain-wind interaction is *orographic* blocking, or simply blocking. Essentially what occurs is that a mountain or mountain range acts as a barrier to low and mid-level winds. In many cases, low-level (roughly within 1 km or 3,280 ft of the ground) air is not capable of being lifted over the top of the range. As a result either the low-level air moves around the sides of the range or it slows down at the base of the mountains forming a pool of air. This does not mean however that all the air on the windward side of the range is blocked. As Figure 4.2 indicates, mid-level air is often able to flow over the top of the blocked layer as well as the mountain range.

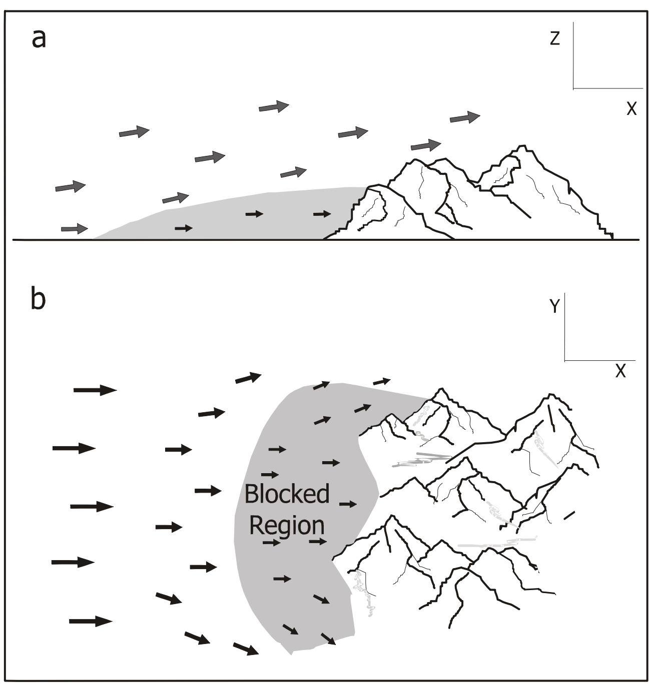


Figure 4.2- a) vertical cross-section of a blocked layer. B) Plan-view. The length of the arrows indicates relative wind speed.

The upstream extent to which blocking occurs is a product of several factors: the stability (change in temperature with height) of the lower troposphere, the height of the mountains, and the speed of the approaching winds. Furthermore if the troposphere is stable and the air is unsaturated, as it usually is, air will cool at the dry adiabatic lapse rate as it rises up the windward slopes of the range. This produces a pool of cooler air at the base of the mountains, which is referred to as a 'blocked layer'. Air within the blocked layer is usually only a fraction of its upstream speed, and its direction is often altered from the original inflow direction as well. In cases of very strong blocking, air within the blocked layer can reverse direction and flow back upstream. The most favorable situation for blocking to occur is when we have a very stable lapse rate (it takes a lot of lift to get air over the mountains), a high mountain range, and weaker winds moving directly toward the range.

If you happen to be hiking on the windward slopes of a mountain range when blocking is occurring, you should expect winds to be <5 m/s (<10 mph) within this layer. Above the blocked layer however, wind speeds return to their upstream values. If you were to hike through the upper regions of a blocked layer, you will experience a sharp increase in wind speed as you ascend.

You may be wondering what happens to the flow as it moves to lee-side of the mountain range depicted in Figure 4.2? Air that has moved around the edges of the barrier can often curve towards the backside of the mountain in what are called leeside vortices. Air that has gone over the top of the barrier can descend to the surface in smooth layers, or if the lee slope is steep, the air often becomes turbulent, forming rolls and eddies. In rather special cases, the flow accelerates as it descends the lee-slope creating a downslope windstorms. Many of these features will be discussed in the following sections.

As you read through the text and view the drawings you should keep in mind that what is being described are idealized scenarios. In real mountainous terrain we rarely find nice symmetrical mountains, instead we find a series of irregularly spaced ridges interspersed with jagged mountains. All of this adds up to produce some very complex air flow patterns.

Barrier Jets

One of the consequences of blocking is the formation of a zone of strong mountain parallel low-level wind called a barrier jet. Barrier jets do not form during every blocking event, in fact well developed barrier jets may only develop several times per year in a given mountain range. In order to understand how this wind develops, imagine a layer of air which is moving perpendicular to a mountain range, at some height above the ground as drawn in Figure 4.3 As a result of blocking, a mesoscale area of higher pressure (mesohigh) forms at the base of the windward slopes (essentially a cold pool of air). Keep in mind that this is a separate feature from the synoptic-scale pressure gradient which generates the winds in the first place.

Now the speed and direction of the flow upstream of the range is a balance of the pressure gradient and Coriolis forces as well as friction. The incoming flow slows down due to the local increase in pressure. Since the Coriolis force is proportional to the speed of the wind, as the wind decelerates in the blocked zone, the wind turns toward the direction of low pressure, which is to the left in Figure 4.3. However, the winds begin to re-accelerate because of increase in pressure that occurs in the cool pool. Even though the Coriolis force tries to shift the winds back towards the right, the mountain range is in the way, as a result the wind remains parallel to the mountains.

The formation of a barrier jet typically takes about 4-6 hours from the onset of blocking. They are fairly common along the west slopes of the Sierra Nevada and along the eastern Rockies. They are much more common during the winter when the lower troposphere is stable. Winds speeds in these jets vary considerably, in very strong cases speeds on the order of 15-30 m/s (35-70 mph) have been reported (Parish 1982). Barrier jets can extend out from the mountains as much as 50-100 km (30-60 miles).

Flow Over and Around Large Mountains

We will now consider what takes place when air moves over and around a large mountain or mountain range. If the summit lies well above the blocked flow or if no blocking occurs, and if the troposphere has a stable lapse rate, wind speeds across the summit will often increase from their upstream free atmospheric values (in other words the winds are accelerated). In order to understand why this occurs, we are forced to consider the propagation of *gravity waves*. Simply put, gravity waves are regions within the troposphere where the air oscillates up and down, in similar fashion to swells on the surface of the ocean. Gravity waves are created because air is forced up and over mountainous terrain, however they only occur when the troposphere is stable (Figure 4.4). Once gravity waves are produced, wave energy propagates away from the mountain in all directions. As a result, the path of the air moving over and downstream of a large mountain or mountain range often has a wave-like structure.

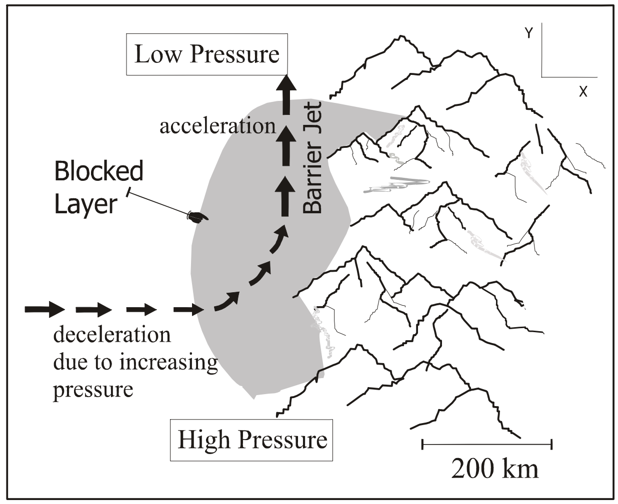


Figure 4.3- Example of a barrier jet.

There are times when the amplitude of gravity waves becomes so large that they overturn or break, like a ocean wave steepening and breaking as it moves onshore. When wave breaking occurs it disrupts the horizontal flow by forming a turbulent region within the middle troposphere. As a result, subsequent air that flows over the mountain is effectively "squeezed" through the space between the summit and the turbulent layer above (Figure 4.4c). Since the volume of air passing between the summit and turbulent region remains unchanged, the flow must accelerate as it crosses the summit and moves down the lee-slope. Wave breaking is not the only way that air is accelerated over the summit of a large mountain, an inversion located several kilometers above the summit, will have the same effect. In this case air is squeezed through the space created by the inversion and the mountain below.

The type of flow described in the preceding paragraph and illustrated in Figure 4.4c is an extreme case. Nevertheless, anytime the middle

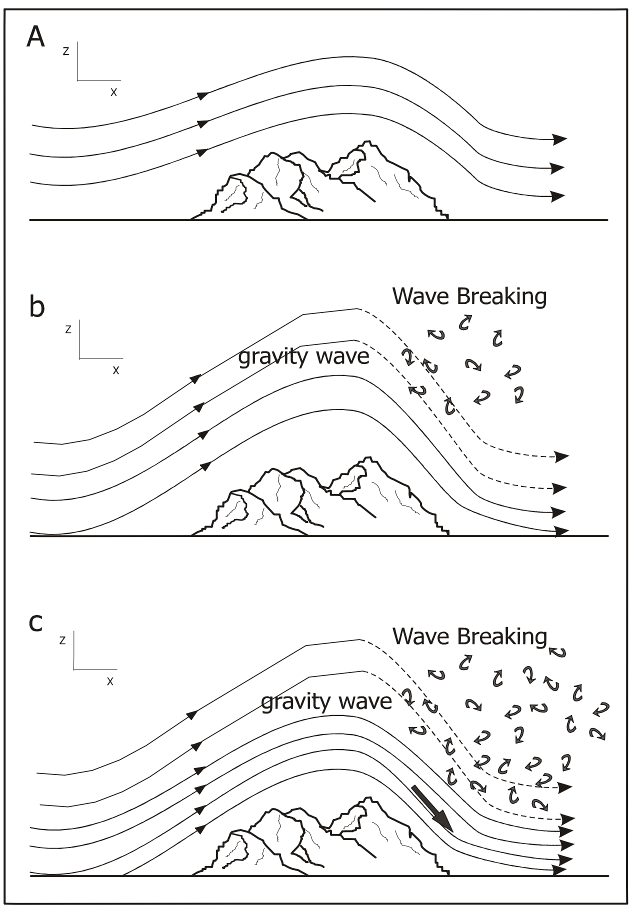


Figure 4.4- Evolution of gravity waves and wave breaking.

and upper troposphere is highly stable, gravity wave propagation above a mountain can cause the upstream winds to be accelerated as they move over the mountain. It would be instructive to compare upstream free atmosphere winds speeds with speeds across the summit of a large mountain. However, since observational data of this nature is rare, a computer model of tropospheric flow was used for analysis. Using model terrain that ranges in height from 2,000-4,000 m (6,660-13,200 ft), and starting the model with a broad range of initial wind speeds, the results show that winds accelerate from 30% and 80% as they move over the summit.

We also need to consider under what tropospheric conditions there is little or no amplification of the winds over the top of a mountain. Table 4.1 highlights five flow regimes where amplification of the wind is minimal.

Table 4.1 Flow regimes with little summit-level wind amplification.

Atmospheric State	Explanation
Weak upstream flow	If upstream wind speeds are below some critical value (roughly 10 m/s or 20 mph), gravity wave development above the mountain is weak.
Unstable mid/upper troposphere	Gravity waves cannot form or propagate in a region where the lapse rate is unstable.
Inversion located at or below summit level	This limits the amount of air that can flow *over* the mountain from low-levels. In addition, at inversion level the winds are often quite weak.
Deep low-level blocking	If the depth of a block layer extends to near summit level, a considerable amount of the flow will be diverted around the mountain, speeds may also be reduced.
Upstream obstacles	If an upstream mountain or ridge generates turbulence or flow deflection, wind amplification will usually not occur over the second summit. This is highly dependent on terrain configuration (distance between ridges), the height of each summit with respect to each other, etc.

In addition to flow over the summit of a mountain, a considerable amount of air can flow around the sides. Additional computer modeling work suggests that in cases of extreme blocking, where the majority of the air approaching the mountains flows around the sides of the barrier, winds on the flank of a mountain can also experience large accelerations (Olafsson & Bougeault 1996). Wind amplification, whether it occurs near the summit or on the flanks of a mountain, is constantly changing as the inflow direction, stability, and upstream wind speed change.

Gap and Gorge Winds

If you have ever been hiking through a mountain gap or pass where the winds were considerably stronger in the gap, when compared to the area just outside of the gap, then you have first hand experience of gap winds. These are the classic type of winds that most mountain travelers envision when meteorologist talk about the acceleration of winds in an mountainous environment. In this book we will refer to a gap as a small elevated divide on a ridge or as a pass through the mountains; where the width is roughly equal to the length. Winds that are funneled through gorges and valleys, will be referred to as gorge winds. [Note that the term valley wind is reserved for a different class of wind discussed later].

Let's start this discussion by considering an example from hydraulics that will serve as an analogue to gap and gorge winds. Many readers are familiar with the "venturi effect", in which a fluid accelerates as it is forced through a constriction in a pipe or hose. By way of illustration consider water flowing through a solid piece of pipe that is 10 cm (4 in) in diameter as depicted in Figure 4.5. What happens to the flow of water if the end of the pipe narrows down to 5 cm (2 in)? If the pressure on the upstream end of the pipe is constant, then water is forced to accelerate through the constriction. Basic physics shows that the speed of the water in the 5 cm diameter section is roughly 4-times faster than in the 10 cm (4 in) diameter section. Most of the acceleration occurs in the transition section of the pipe.

In the example cited above, the constriction in the pipe plays the role of a mountain gap. As air flows through a gap it is accelerated because higher pressure forms near the gap entrance. Gap winds are generally not 4-times faster than the upstream wind speed because of other effects not discussed, nevertheless gap winds can be quite substantial.

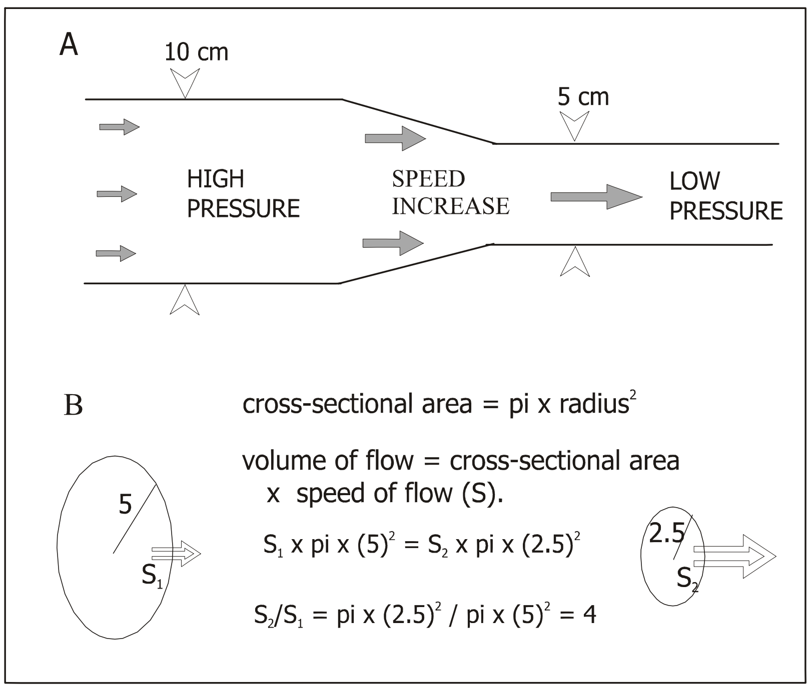


Figure 4.5- Flow of water through a pipe.

The example given in Figure 4.5 is an 'instantaneous' snap-shot of the flow, but what happens if we observe the flow over a period of time? In essence the pressure in the pipe or gap is not steady in time, this in turn produces surges in the flow. For example, in Figure 4.5, if the volume of the water passing through the constriction is smaller than the volume of water which enters the pipe on the upstream end, then the pressure in the wide section of the pipe is forced to increase. This increase in pressure in turn produces an extra acceleration in the flow through the constriction. This 'tug-of-war' between water pressure and the speed of the flow is what generates the surges of water in the pipe. If we extrapolate this example to air moving through a mountain pass, then it is not difficult to understand why the winds are often gusty.

Gorge winds are similar in nature to gap winds in many respects, however, due to the greater length of gorges, the flow is not uniform down the long axis of the channel. In areas where the gorge narrows, flow usually accelerates while in wide reaches it often decelerates (Figure 4.6). It is also fairly common to find hydraulic jumps within a gorge flow. Hydraulic jumps occur when the wind speed decreases and the depth increases over very short distances, resulting in a wave-like feature or abrupt 'jump' in the depth of the layer of air. Hydraulic jumps form in rivers as well and are generally avoided by river runners since large jumps are highly turbulent.

Some of the more notable gorge winds are found in the Pacific Northwest, especially in the vicinity of Fraser Gap and Columbia Gorge. These gorge winds develop during the winter months when a ridge of high pressure produces cold arctic air over southern British Columbia and eastern Washington and Oregon. Jackson and Steyn (1994) found that the strongest winds are within the first 500 m (1650 ft) of the surface and that the depth of the cold air decreases down the long axis of the gorge. The cold air in a gorge is often bounded by an inversion, which in this case acts like a rigid lid forcing the flow beneath it to accelerate. At the outlet of gorges, wind speeds are often two or three times what the ambient pressure gradient alone would have generated. Typical speeds of 20-30 m/s (44-66 mph) were reported by Mass *et al* (1995) for Fraser Gap winds, with some observations of gusts as high as 45 m/s (100 mph) over the San Juan Islands.

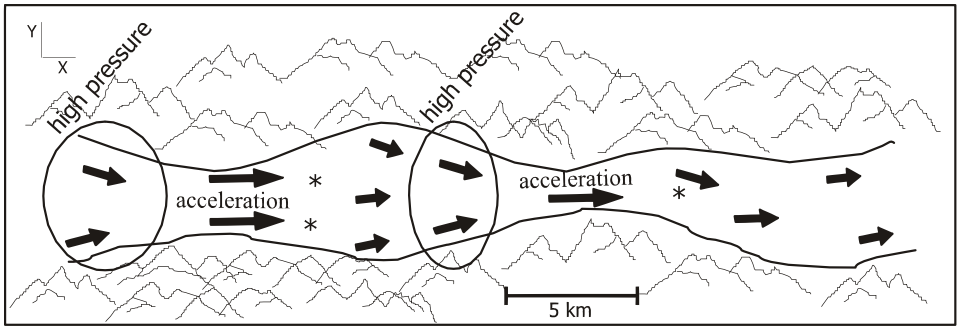


Figure 4.6- Gorge winds. In locations marked with a * it is common to find hydraulic jumps- areas where the wind rapidly decelerates.

Air Flow over a Ridge (microscale)

Understanding how air moves over a ridge should be of particular interest to climbers, skiers, and snowboarders; since it has important implications on the distribution of snow. This section should help explain why windward slopes often have a thin snowcover, while leeward slopes are buried in snow.

As air moves over a ridge, it usually reaches its maximum speed at the crest of the ridge, as shown in Figure 4.7. Essentially this case is no different than the gap winds described in the previous section. In this new scenario however, the 'gap' is formed by the ridge itself and the stable layers of air some distance above the ridge. This means that air which is moving up the windward slope is forced to converge at the summit. To the lee of the summit air moves into a region where the flow is divergent (i.e. expands), hence the speed decreases. Furthermore as a result of the decrease in speed, a localized area of high pressure forms to the lee of the ridge. If the angle of the leeward slope exceeds 15^{o}-20^{o}, then air moving down the slope no longer remains in contact with the surface.

At times air is redirected back up the slope in the form of a lee-side eddy, due to the presence of the localized area of high pressure. These eddies often contain significant amounts of turbulent air, nevertheless their speeds are a fraction of the wind speed in the free atmosphere above. This whole process is referred to as flow separation, which reflects the fact that a fairly homogenous flow regime develops into two different regimes because of the steep leeward slope.

We will take a closer look at the distribution of snow by wind at the conclusion of this chapter, however this conceptual model of flow over a ridge helps explains why snow is frequently transported from windward slopes, where wind speeds are high, to the leeward slopes, where speeds are much slower. In fact the amount of snow that the wind can transport is a function of the speed of the wind. High winds can transport large amounts of snow, once the speed decreases however, the carrying capacity of the wind decreases as well. As a result, large quantities of snow are typically deposited on leeward slopes in areas where wind speeds diminish and eddies form. The complicating factor is that the wind does not always blow from the same direction. For example, snow can also be deposited in gullies and couloirs when the wind blows parallel to the long-axis of a ridge, in a process called cross loading. In this case the upstream side of the gully plays the role of a ridge, snow is deposited at the base of the upstream slope as wind speeds start to decrease.

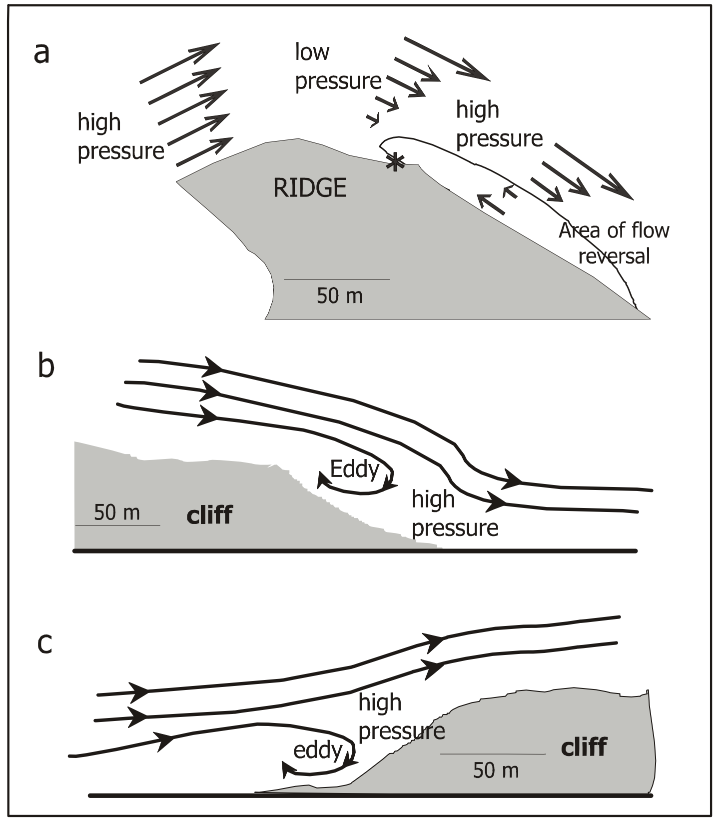


Figure 4.7- a) flow over a small ridge. b) flow over a cliff. c) flow up a steep slope.

Flow separation can occur over a variety of topographic features as illustrated in
Figure 4.7 b,c. As air encounters a steep cliff or mountain face, the air decelerates at the base of the cliff and forms a eddy due to an increase in local air pressure at the base of the cliff. Flow separation also explains why wind breaks, like a small grove of trees or a outcrop of rocks, work as well as they do. As a rule of thumb, wind breaks are effective to a distance of about 2 or 3 times their height. For example, if a group of trees or rocks are 10 m (33 ft) high, the zone of reduced winds extends out to a distance of about 20-30 m (65-95 ft), as measured from the base of the barrier.

Downslope Winds

These types of winds occur when air flowing over a mountain barrier accelerates on its descent down the lee-slope, as illustrated in Figure 4.8. Downslope windstorms occur in virtually all moderate to large mountain ranges and are much more common during the cooler months of the year than during the warm months. They are also associated with rapid increases in lee-side temperatures, leading to rapid snow melt. Downslope windstorms are commonly know in North America as *Chinooks* and in Europe as Foehns.

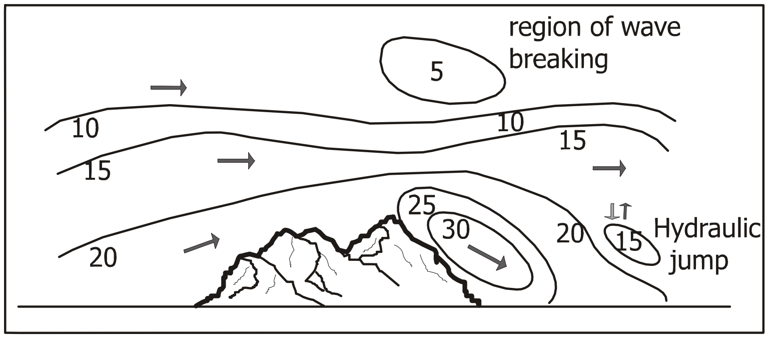


Figure 4.8- Downslope windstorm with contours of windspeed (isotachs) in ms^{-1}.

In North America downslope windstorms are well documented, large events occur on average several times per year in the western USA. They are most common in the Front Range of the Colorado, along the east side of Glacier and Waterton Lakes National Parks, and in the Canadian Rockies, just east of Banff-Jasper. They are also fairly common on the east side of the Sierra Nevada's, near Bishop. In Europe, foehns are common on both the north and south side of the Alps, depending on the direction of the geostrophic winds. When downslope winds occur, strong gap and gorge type winds may also be produced as air flows out of the mountain passes and valleys.

In order for a downslope windstorm to develop, the following requirements need to be meet (Brinkmann 1974): 1) A mountain range has to be high enough and large enough to disrupt the air flow over its crest. Isolated mountains (like stratovolcanos), even large ones are not prime candidates for downslope windstorms because air often flows around the sides instead of over the summit. 2) There must be moderate-to-strong winds (>20 m/s or 44 mph) at mountain crest level.
3) The direction of the wind should be nearly perpendicular to the long axis of the barrier. 4) The air over the mountain must have a stable lapse rate. The steeper the leeward slope, the easier it is to generate strong downslope winds. However, the steepness of the windward slope is not important.

The January 11-12, 1972 downslope windstorm in Boulder, Colorado (Klemp & Lilly 1978) was noteworthy because it produced a considerable amount of damage in the area. In this particular storm wind speeds of 50 m/s (110 mph) were observed at ground level near the base of the Front Range. One common feature of Front Range wind storms is the presence of an inversion, roughly 2 km (1.2 miles) above the highest terrain. When a downslope windstorm does occur, the area of strong winds usually does not extend much more than 30-50 km (20-30 miles) from the base of the mountains. In addition, it is common for the downstream end of the strongest winds to terminate abruptly in a hydraulic jump.

As you can see from this short description, a number of tropospheric variables, primarily the stability, wind speed and wind direction have to be 'in alignment', before a downslope windstorm can occur. It turns out that this alignment occurs less frequently than you might imagine. As a result, areas known for downslope winds might typically experience three to five significant events and 10 to 15 minor events during the course of a year. Of course there are many mountainous regions around the globe where these events go unnoticed because there are no systematic weather observations.

Discussion: As a mountain traveler take note that the strongest winds in downslope windstorms extend from about halfway down the lee slope to the base of the slope. The good news is that most downslope windstorms last less than 12 hours. They can occur in clear air or in association with clouds. In many cases clouds and precipitation are produced on the windward side of the barrier and over the crest.

On the lee-slope however, air is drying as it descends, so partly cloudy skies are the norm. The leeward temperatures are generally warmer than the corresponding temperatures at the same elevation on the windward slopes. The reason for this is that due to the release of latent heat as moist air ascends the windward slopes, the air over mountain range is relatively warm. As the air moves down the leeward slope it is compressed and hence it warms even further.

If you are in the mountains and you think you are in a situation where a downslope wind event is beginning, you will first notice an increase in wind speed and rising temperatures. There are no real precursors or atmospheric parameters that you can monitor from the field that will suggest a downslope wind event is going to develop. About the only thing you can do is find a sheltered location and wait it out.

There are a few places in the world where downslope winds lead to a cooling on the lee-side of the range, these events are called boras and are common in the Dinaric Alps (western Slovenia and Croatia), as well as the region between the Pyrenees and the Alps (Smith 1979). Boras from when a deep layer of cold air accumulates on the windward side of the mountains. As air flows over and down the lee side of the range it of course warms . However, what warming does occur is not enough to off-set the initially cold air temperature. Therefore at the base of the leeward slope, temperatures become cooler instead of warmer. Boras occur in ranges where the mountains are of modest height. If the mountains are too high, cold air can never become deep enough to spill over the top, although some air will flow through gaps and gorges producing localized areas of cold air drainage.

Some of the diagrams in this section may have given the reader the impression that downslope winds only occur on the leeward slope of an isolated mountain range. In reality, they can occur to the lee of any range that meets the stated criteria. It is possible to be in the middle of a large mountain range and still experience downslope winds. As was noted earlier in this chapter, the type of flow that occurs in the middle of a mountain range depends on the distance between the mountains and their relative heights. In addition, if the intervening valleys are full of cold air it is unlikely that a downslope windstorm will develop, even if all other factors are favorable.

Lee-Waves and Rotors

There is a special class of mountain induced gravity wave phenomena called lee-waves which are characterized by a series of troughs and crests in the flow directly above and downstream of a mountain barrier (Figure 4.9). This type of flow has much in common with downslope wind storms, however there are some significant differences as well. When flow over a mountain generates lee-waves, it means that wind speeds near the summit and above are quite high. However, unlike downslope windstorms, surface winds on the leeward slope and at the base of the mountain are usually pretty light. The generation of lee-waves depends on the vertical profiles of wind speed and atmospheric stability as well as the height

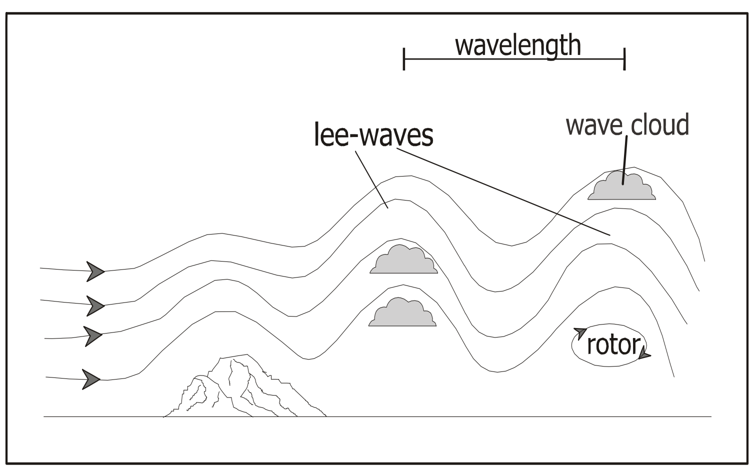


Figure 4.9- Lee-waves with wave clouds and rotor.

and width of the mountain.

Lee-waves stay fixed with respect to objects on the ground even when strong winds are blowing through the waves. This occurs because lee-waves are a collection of gravity waves that have their own individual characteristics. When these gravity waves are combined to produce lee-waves, they form a stationary pattern in the air above the mountain. The height of lee-waves is strictly a function of mountain width and height; where a narrower and steeper mountain produces higher and thicker lee-waves.

The wind speed and stability on the other hand, determines the horizontal wavelength of the lee waves (distance between successive waves). Durran (1986) states that most lee-waves have horizontal wavelengths between 5-25 km (3-15 mi) and that the minimum speed of summit-level winds needed to generate them is on the order of 10-15 m/s (25-35 mph). It is quite common to have a wave form directly above the summit of a large mountain as well. If sufficient moisture is available in the middle troposphere, clouds may form in the crest of these waves (lenticular clouds), which is the only way an observer on the ground knows of their presence. A pilot on the other hand may experience moderate to heavy turbulence when flying through or near lee-waves.

Rotors form on the downstream side of mountain ranges when mid-tropospheric winds form an eddy-like feature. Rotors are like a continuous series of ocean waves, that break in one specific spot as they move toward the beach. They most often form in conjunction with lee-waves or downslope windstorms. They may reach the ground for short periods of time but more often remain aloft. Clouds usually do not form in rotors, although if a rotor is located near the ground it can stir up large quantities of dirt, and be a serious hazard to aircraft.

Wind Flow over a Valley

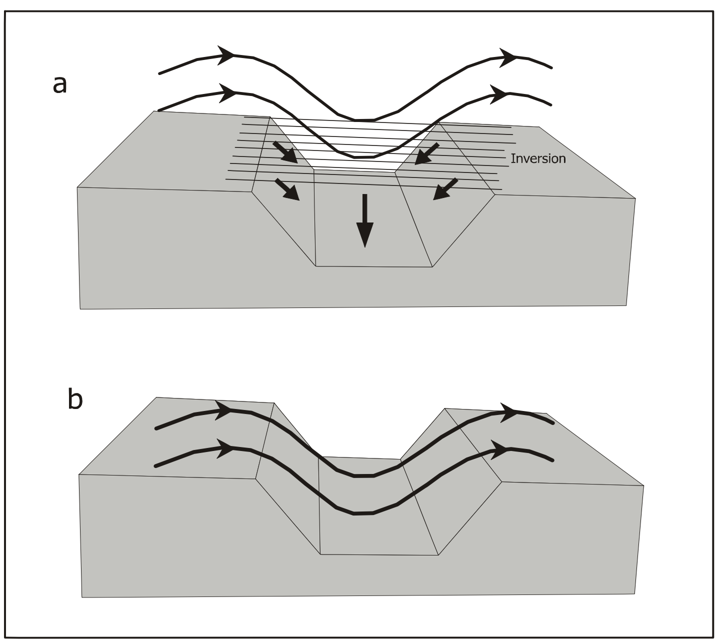


Figure 4.10- a) Due to an inversion, geostrophic winds cannot enter the valley. b) Without an inversion winds descend deep into the valley.

We have already discussed air flowing down the long-axis of a valley, but what happens when synoptic-scale winds blow over the top of, and perpendicular to the long-axis of a valley? Figure 4.10 shows several possible flow scenarios. Not all larger-scale winds descend into the valley, in fact they often do not. The determining factors are:

* Valley width
* The height of terrain upstream of the valley
* Stability of the air in the valley
* Presence of an inversion at the top of the valley
* Strength and depth of thermally generated winds within the valley
* Speed of the large-scale (geostrophic)

winds

Thermally Generated Winds

This class of winds should be of great interest to mountain travelers, because they have a high frequency of occurrence and are found in every type of mountainous environment. Since all winds are ultimately thermally generated, you are probably wondering what specific type of winds we are talking about? Thermally generated winds are essentially near surface winds that develop in response to differential heating/cooling of the earth's surface. You may recall from chapter 3, that heating of the ground by shortwave radiation and the subsequent cooling by longwave radiation occurs on the local-scale, basically from slope-to-slope. This is why sun angle, slope aspect, latitude, time of year, and physiographic properties of the surface determine when and where thermally generated winds will form. These types of winds usually only develop when the low-level geostrophic winds are light (< 5 m/s or 10 mph).

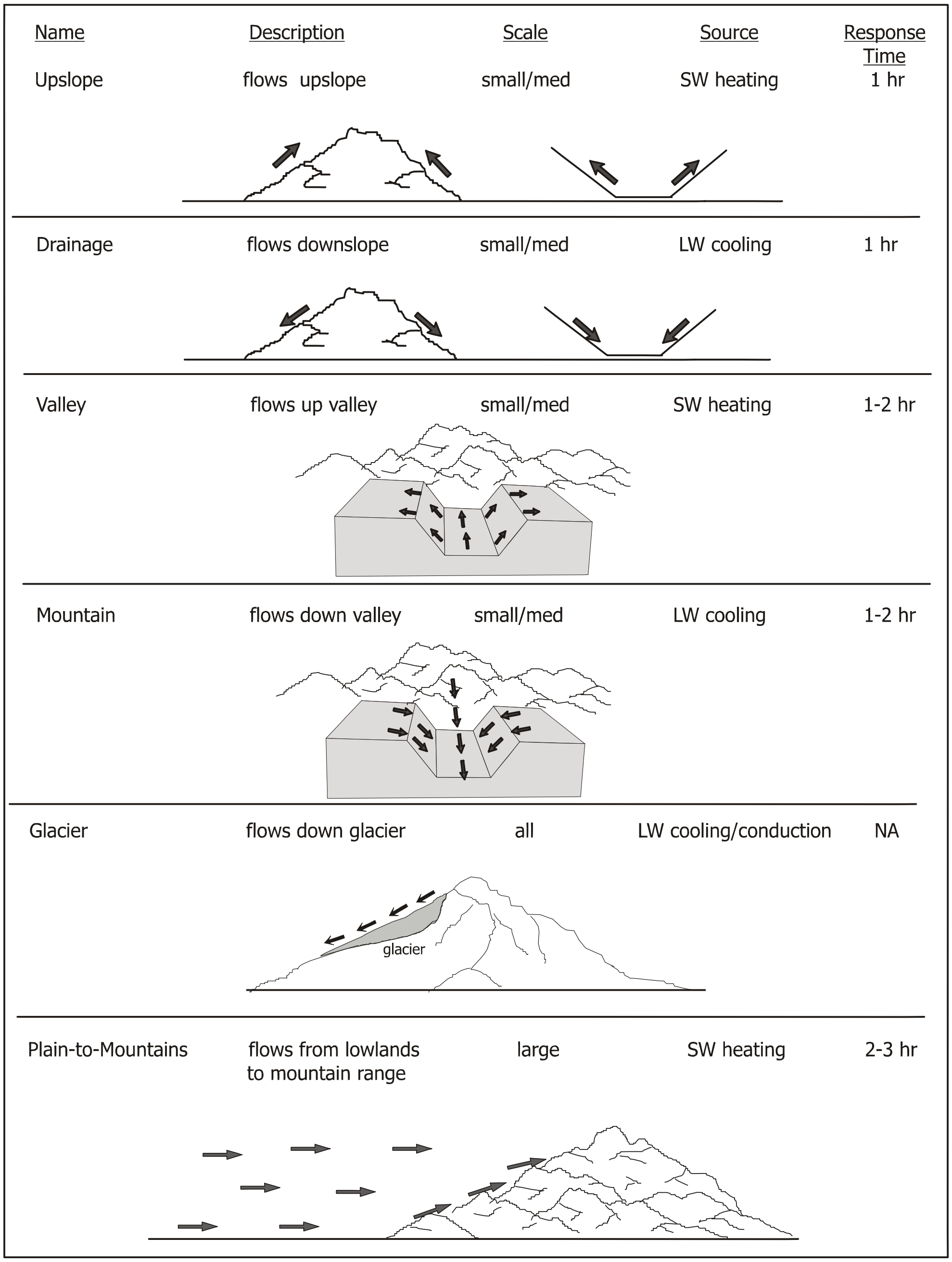


Figure 4.11- Various examples of thermal winds.

Figure 4.11 illustrates the various types of thermally generate winds which will be discussed in this section. In addition, this figure also gives an overview on the spatial scale at which they occur, what process(es) are responsible for their development, and the amount of time it takes for them to develop.

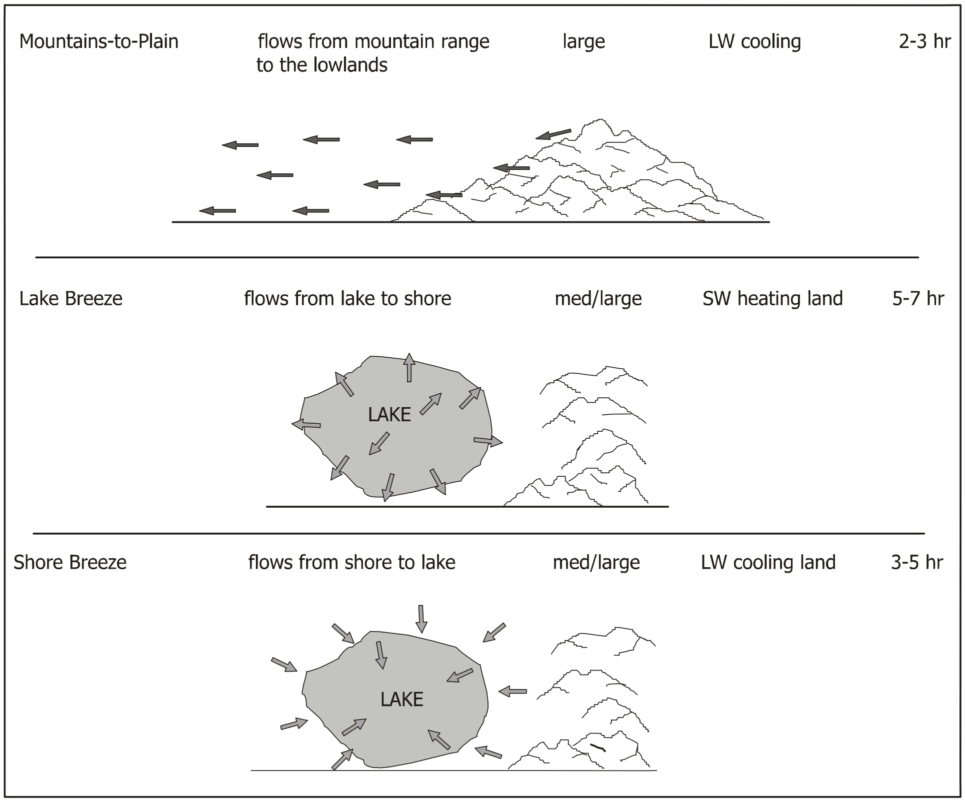


Figure 4.11 continued

As a rule of thumb, small-scale winds (those that involve a smaller volume of air) develop on the order of 30-60 minutes, while large-scale winds take 1-3 hours. For example, in small alpine valleys, upslope and drainage winds can begin to develop 20-30 minutes after sunrise or sunset, respectively. On the much larger plain-to-mountain scale, thermally generated winds may take several hours to develop since considerably more air has to be heated. Glacier winds on the other hand develop independently of the solar cycle, and can blow for days on end if the geostrophic winds remain light.

Slope, Mountain-Valley, Mountain-Plain Winds

In order to understand how these winds develop, consider the idealized two-dimensional valley configuration in Figure 4.12. We will assume there are no ambient winds, the ground is snow-free, and clouds are not a factor. Figure 4.12a shows what the thermally generated wind structure might look like in the early morning, shortly after sunrise. The air next to the valley floor and slopes heats up because of its contact with the rapidly warming ground. The air in the middle of the valley on the other hand is considerably cooler than the air near the sunlit slopes or valley floor. Since warm air is less dense than cooler air, it rises. Slopes that are in the shade may still have cold air moving downslope at this time, but by late morning, upslope winds are usually fully developed throughout the entire valley. At night, because of extensive longwave cooling of the surface, the air nearest the ground cools more than the air in the middle of the valley atmosphere (Figure 4.12b). This results in air moving down the slope in the form of a drainage wind. Studies have shown that the speed of a drainage flow is proportional to the steepness of the slope, as well as the temperature gradient between the air near the ground and the air at the same height but some distance out from the slope. The larger the temperature gradient and the steeper the slope, the faster the flow.

Both upslope and drainage winds have typical depths on the order of 10-30 m (30-100 ft), and speeds from 2-5 m/s (5-12 mph). From time-to-time however both the depth and speed of these winds can be much larger. You should note that upslope and drainage winds develop above many types of sloping terrain, it does not matter if the sloping terrain is on the flanks of a mountain or the sides of a valley. What does matter to a large degree is the type of ground cover- a wet heavily vegetated slope will take a lot longer to warm up then a dry rocky slope. As we will see below, a slope covered by snow or ice will not heat up at all. It the sky is overcast, these winds may not develop at all.

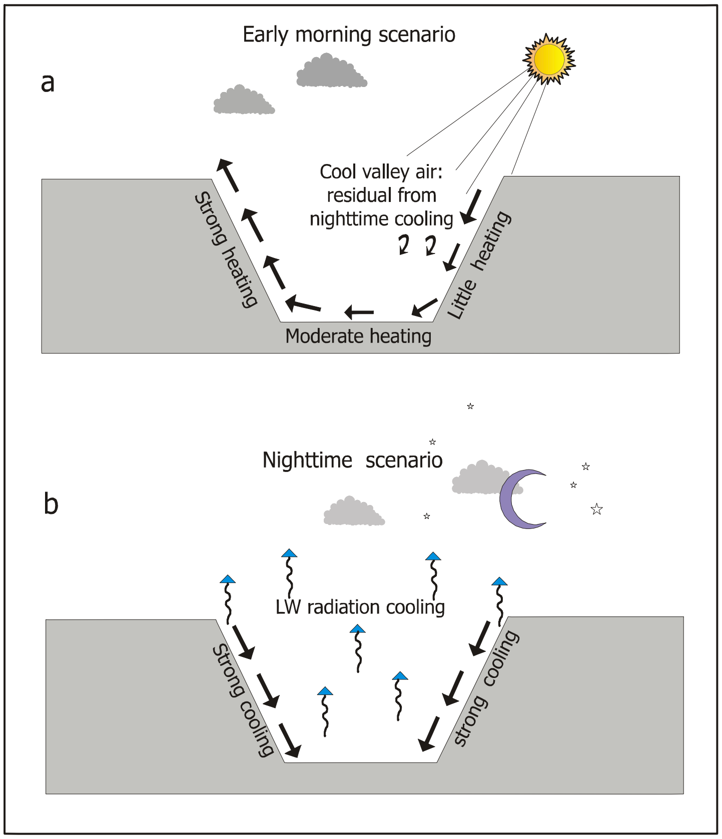


Figure 4.12- a) Generation of upslope flow in the morning.
b) Drainage flow at night.

The next step is to consider winds that form parallel to the long-axis of a valley. During the course of a day, air within a valley heats up considerably more than the air at the same elevation outside of the valley over the lowlands. This produces a valley wind that moves up the valley towards the mountain (Figure 4.13). At night, since air in the valley cools more than air over the lowlands, the valley air becomes cooler and more dense. The resulting flow moves down valley in the form of a mountain wind. These types of winds have a depth on the order of 50-200 m (160-650 ft) for typical high alpine valleys, and speeds on the order of 4-7 m/s (10-15 mph).

In large valleys these winds can be much deeper, for example, Vergeiner & Dreiseitl (1987) report valley winds 1-2 km deep in the Inn Valley of Austria. Since valley and mountain winds involve a larger volume of air than either slope or drainage winds, they typically require several hours of heating or cooling of the valley atmosphere before these winds are initiated. In addition, numerous observers have reported finding a layer of winds located just above valley or mountain winds. These winds move in the opposite direction as the valley or mountain wind, and therefore are called anti-valley or anti-mountain winds.

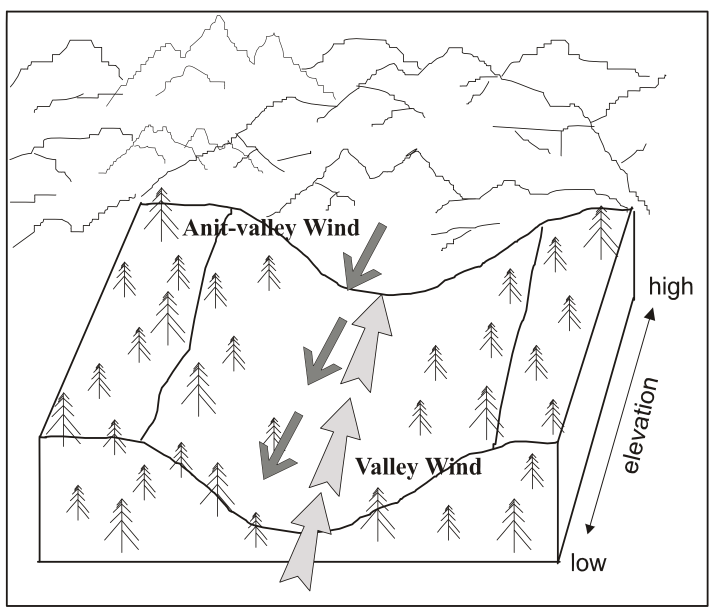


Figure 4.13- Valley wind with anti-valley wind above (return flow).

On a larger scale, a plain-to-mountain circulation can be established during the daylight hours, which reverses direction at night. This type of circulation system is basically a larger scale version of the mountain/valley winds. During the day, the air adjacent to the mountain slopes becomes considerably warmer than air at the same height over the plain, which creates low pressure over the mountains and draws in air from the plains. At night, air over the mountains becomes cooler than air over the plains, so cold air flows down from the mountains. Plain-to-mountain winds generally develop three to four hours after sunrise, while mountain-to-plain winds develop an equal number of hours after sunset.

What we have been discussing so far in this section are idealized models of how thermal winds develop. In the real world they tend to be much less homogeneous in space and steady in time. Added complications arise due to the non-homogeneous nature of terrain and physiographic features. For example, variations in slope angles, ground moisture, vegetation types, snowcover, etc. all have important effects on thermally generated winds. Further complications come from the development of inversions during the night, the interaction of geostrophic winds, and cloud development.

The following example illustrates that our conceptual models are essentially correct, however real world flows tend to be much more complex. Case in point, the world's deepest valley, the Kali-Gandaki of Western Nepal (located between Dhaulagiri and Annapurna massifs), was the site of two short field programs that studied the notorious local mountain-valley winds, during the summer and early autumn seasons. Researchers at one point used a motorized glider to monitor these winds (Neininger *et al* 1985, Egger *et al* 2000). Near the valley floor they measured valley winds that had sustained speeds of 10-15 m/s (20-35 mph) with gusts as high as 25 m/s (55 mph). Wind speeds were not uniform through the length of the valley, they tended to be highest in the central section.

At times valley winds would persist through the night, albeit in a much weaker state. On other nights however, the valley wind would be replaced by a weak mountain wind. The high speeds of the valley wind are attributed to the massive heating which takes place on the Tibetan Plateau during the day. The heating of the surface of the plateau creates low pressure over Tibet, which draws higher pressure air located over central Nepal, through the valley into Tibet. This is a prime example of a mesoscale plain-to-mountain flow combining with a valley wind to generate exceptionally persistent and strong winds.

During a separate field program, Ohata *et al* (1981) spent an entire year observing thermally generated winds in a unnamed valley of the Khumbu Himal region of eastern Nepal. One curious phenomenon they observed during the monsoon season (June-September) was the persistence of valley winds long after sunset. There was significant cloud formation in the late afternoon and early evening which caused the valley atmosphere to remain warmer due to the release of latent heat, than the time of day alone would suggest. Their data also revealed a weaker thermal wind regime during the monsoon season, compared to the pre- and post-monsoon seasons. Weaker winds during the monsoon season were attributed to an increase in surface moisture at low elevations, greater snow cover at higher elevations, and greater cloud coverage.

Buettner & Thyer (1966) spent several summers studying thermally generated winds in the 5 km (3 mi) long Carbon River Valley located on the northwest flanks of Mt. Rainier. They observed that the typical depth of a valley/mountain wind was from one-quarter to one-third the depth of the valley. The overlying anti-wind had about the same depth. Valley winds reached maximum velocity in early afternoon while mountain winds peaked just before sunrise. Typical speeds for both wind regimes was on the order of 3-6 m/s (7-13 mph). They also noted that speeds were rarely constant and seemed to vary with a period of about 20 minutes.

Glacier Winds

When a stagnant layer of air lies over a large mass of snow or ice, such as a glacier, there is considerable cooling of the air due to conductive heat loss to the ice below. In time this layer of air becomes significantly colder than the ambient air at the same elevation in the free atmosphere. As a result, cold air moves down slope in the form of a glacier wind. Like all other types of thermally generated winds, glacier winds usually only form when the geostrophic winds are light. Glacier-type winds also develop over snowfields (snow breezes), the air adjacent to the surface does not care whether it is in contact with snow or ice. Glacier winds and snow breezes can also develop over horizontal ground, as long as a large temperature contrast is maintained between the glacier (snow) and the non-glaciated (snow-free) ground.

Unlike other thermal winds which often change direction at sunrise and sunset, glacier winds and snow breezes are generated independently of the solar cycle. This should not be taken to mean that glacier winds maintain constant speeds. In fact most of the glacier winds that have been studied display some type of diurnal wind speed fluctuation. The speed of a typical glacier wind ranges from 5-10 m/s (12-24 mph), with a depth generally not exceeding 100-200 m (300-600 ft). Speeds are not constant with height, there is usually a distinct speed maximum, which can occur however, at just about any height. Due to the partial dependency on longwave cooling of the air adjacent to the ice/snow, glacier winds tend to be fully developed on days/nights which are predominately cloud free.

Over the past 40 years there have been a considerable number of studies on the subject of glacier winds. What all these studies show is that there is significant variation from one glacier to the next. This testifies to the wide range of conditions and spatial scales under which glacier winds have been studied, and more importantly it tells us that we cannot assume that all glacier winds have identical properties.

Glacier winds, like all thermally generated winds, exhibit speed fluctuations due to turbulence. They can also extend well beyond the terminus of the glacier. During the day, a glacier wind may flow down valley and converge with a valley wind, creating a zone of vertical motion that is a preferred area for cumulus cloud formation. At night, glacier winds can reinforce mountain and other drainage winds as they all move down valley. In general, glacier winds will reach maximum speeds on or near the terminus, since beyond the terminus they encounter an increase in surface roughness (trees, large boulders, etc.), which decelerates the flow. As the studies cited in this section indicate, glacier winds show considerable variation in depth, speed, and extent. This illustrates the difficultly of trying to construct a conceptual glacier wind model that fits all cases.

Lake Breezes

Since thermally generated winds result from the differential heating of the earth's surface, the heating/cooling contrast between a lake and adjacent land is sufficient to generate a lake-to-land breeze during the day (lake breeze), and a land-to-lake breeze at night (land breeze). The development of these breezes over a large lake is a result of the different thermal properties of water versus land (dirt or rock). As a starting point for this discussion we will assume that the water and the land have the same temperature at sunrise, we will also assume that over the course of the day equal amounts of shortwave radiation is absorbed by the water and dirt (we are neglecting reflection off of the water). As a result, equal amounts of energy are absorbed by each square meter of the respective surfaces. However, if we were to measure the temperature of each surface during the early afternoon, we would find that the land is considerably warmer than the water, why? Water conducts heat much better than dirt or rock, hence heating occurs deeper within the water column.

In addition, since the density of water is temperature dependent, small circulation cells are initiated within the upper few centimeters (inch or less) of the water layer. The net result is that the energy absorbed by the water is spread over a larger volume of material as compared to what occurs in dirt or rock. In fact most of the shortwave radiation absorbed by the land, stays in the uppermost layer (few millimeters), giving it a very warm surface temperature. This has an important effect on the temperature of air overlying each surface. By afternoon the air over the land is warm and the air over the lake is still relatively cool. This results in low pressure developing over the land with slightly higher pressure over the water. This pressure gradient in turns produces a light wind which blows from the lake to the shore.

During the evening and early night time hours, both the water and the land cool. Since the land does not retain very much heat (low thermal heat capacity), it cools rapidly due to emission of longwave radiation. On the lake the upper 10 cm (4 inches) of the water still has a considerable amount of heat in storage, which is by the way slowly released over the course of the night. As the surface of the water cools due to the emission of longwave radiation, heat energy from the water below is conducted to the surface, replenishing the heat loss. Therefore, at night the air over the land is cooler than over the water, and a land-to-water breeze is generated.

Lake (or land) breezes typically range from 2-3 ms^{-1} (4-6 mph) and only occur on days with considerable shortwave radiation input. As you might imagine, on days with extensive cloud cover or when the land is moist, these types of breezes are suppressed or at least delayed.

Wind and Snow

We have already briefly discussed the transport of snow from the windward to leeward side of a ridge, this section will however elaborate on the processes involved. We will use the term snow “particles”to designate snow that is already on the ground or picked up and transported by the wind, and reserve the term snow “crystals” for snow which is falling from clouds as original precipitation.

Movement of Snow by the Wind

There are three ways that snow particles can be transported: rolling, saltation, and suspension (Figure 4.14) In reality the same forces are at work for each of these three mechanisms, the difference being how far above the surface the snow particles are lifted. It is important to consider the degree of bonding between the individual particles. Shortly after being deposited on the ground, snow particles quickly form bonds with adjoining particles. The stronger the bonds and the larger the particles, the more difficult it is for the wind to move them. The degree of bonding is a function of the past and current air temperatures. Cold dry snow is more susceptible to wind transport than wet heavy snow, which has a temperature near freezing. Research has shown that the bonding strength of particles increases as the temperature of the snowpack approaches 0^{o} C (32^{o} F).

Rolling occurs when wind speeds are not strong enough to lift a particle off of the ground. Due to wind stress as air moves over and around a particle, the upstream side of a particle is subject to higher pressure than the leeward side. This is what causes particles to either roll or slide across the surface. Angular particles tend to have sharp corners broken off, so that after a few minutes of rolling, most particles have a rounded shape.

Saltation occurs when the wind speed is large enough to create a small amount of aerodynamic lift, which carries the particle into the air stream. The lift is created in the low pressure zone on the back side of the particle. However, due to its weight, the particle starts to fall back towards the surface, following a parabolic trajectory. When a particle hits the surface, it can loosen surrounding particles and make them more susceptible to lifting, since the impact weakens or destroys ice bonds. Some snow scientists (the folks who get paid to play in the white stuff) suggest that the impact of a falling particle actually works like an elastic collision (the transfer of momentum from the particle that is landing to the one that is being ejected), sending new particles into the air stream. One of the best ways to understand the saltation process is to observe it on a windy day at the beach. If the wind speeds are moderate, dry sand can be lifted above the surface, only to travel a 10-20 cm (5-10 in) before crashing into the surface and ejecting other sand particles into the air stream. There is a minimum wind speed at which the saltation process begins in snow, the exact value of course depends on the amount of bonding.

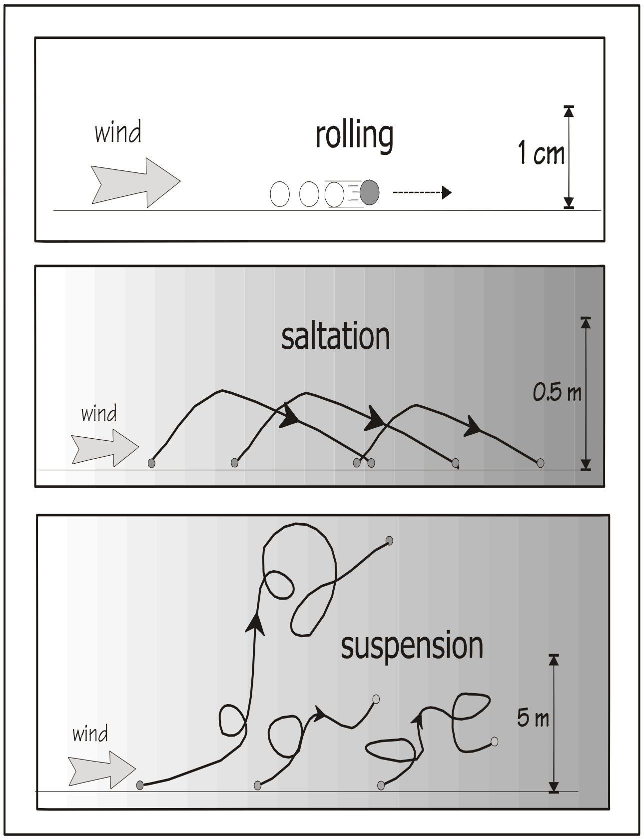


Figure 4.14- Three modes of snow transport.

Mellor (1965) for example, found that speeds from 6-10 m/s (13-22 mph) were high enough to initiate saltation in dry snow. Empirical data also shows that the mass of snow transported increases exponentially with increasing wind speed. The depth of a typical saltation layer is on the order of 0.5-1 m (20-40 in), with a typical horizontal trajectory of less than 2 meters (~ 6 ft).

The third and final way that snow particles are moved is by suspension, also called turbulent diffusion by some authors. When wind speeds are high (approximately > 15 m/s or 32 mph), it is possible for particles to be carried horizontally for 10-100 meters (30-300 ft) without making contact with the surface. Snow that is transported across a ridge undergoes saltation or suspension, or both. The determining factors are wind speed, particle size, and degree of bonding.

By virtue of continued movement, wind transported snow particles become rounded, typical diameters range from 0.1-1.0 mm (0.004-0.04 in). Snow drifts and snow slabs have a density range of 200-400 kg m^{-3} (12-24 lb ft^{-3}), where glacier ice for comparison has a density of 700-900 kg m^{-3} (43-55 lb ft^{-3}) and fresh water 1000 kg m^{-3} (62 lb ft^{-3}). The tight packing of rounded particles in wind slabs results in a layer with considerable mechanical strength (it can take a lot of stress without breaking). However, these dry particles <u>do not</u> bond well with older types of snow that lie underneath or next to the slab. The mechanical strength is a mixed blessing. The slab itself is quite strong, however, pressure waves can travel great distances through it, so when failure does occur, it is generally catastrophic (involves a much larger volume than it would in unconsolidated snow).

As backcountry skiers know all to well, skiing across wind blown snow is often difficult due to the presence of sastrugi and dunes. The steep side of sastrugi point in the direction of the prevailing wind during the time of formation (these are created by the removal of snow). In addition, when skiing from unconsolidated snow to area of wind blown snow, skis often slow down and feel "sticky" because wind blown snow is very dry. Wind blown snow is dry because surface moisture was removed or shed while the particle was being transported.

Cornices form when snow is blown to the crest of a ridge. Snow collects on the very edge and forms a horizontal protrusion. In time, the pre-cornice grows vertically as well as horizontally. Due to the strength of the bonds among the particles, large volumes of snow are able to hang out over the edge of the ridge and not break off. Over the course of the winter, due to its weight, the cornice begins to deform and take on the classic curl shape

---

Discussion: If you have ever camped in a tent during a blizzard, you know that the rate of snow accumulation on and around a tent can be phenomenal. For a given rate of precipitation (the amount of snow falling from the clouds), the amount of snow that accumulates on a tent increases dramatically with increases in wind speed, up to a point. What happens is that the wind blows snow into the sides of a tent at a much faster rate than it would accumulate if the wind was calm and the snow was falling straight down. If the winds are very strong (> 25 ms^{-1} or 55 mph) during a blizzard, then chances are not very good that very much snow will accumulate, because most of it is blown away. One of the keys to successfully enduring a major snowstorm is to be mentally prepared to occasionally leave the tent to clean off of it, or expend the energy to dig a snow cave.

Snow Fences

Marring the landscape as they do so well, snow fences play a key role in helping man attempt to tame nature. Snow fences are more commonly used in the Alps than in any mountain range in North America, in large part because the Alps sustain a much larger population base.

The sole purpose of a snow fence is to control the deposition of snow. There are several different types of snow fences, two of the more common ones are blower fences and collecting fences. Blower fences are designed to accelerate the wind in a given area, so that its snow load will be deposited farther downstream, much like wind accelerating through a mountain gap. Collecting fences on the other hand are used to reduce wind speeds by generating turbulence, so that snow is deposited directly to the lee or around the fence. If the fence can disrupt the flow, wind speeds will decrease, and the carrying capacity of the wind will be reduced. Snow fences are used to reduce the amount of snow accumulating across roads, railroad tracks, buildings, etc., which saves removal costs and reduces potential hazards (Figure 4.15). Collecting fences may also be used to increase the amount of snow accumulation near a lake or pond, in hopes of increasing spring runoff into the lake.

Collecting fences are most often constructed as non-solid barriers (they have open spaces between the slats), made from wood or metal and are from 2-3 m (6-10 ft) in height. Slats can be placed either horizontally or vertically. The height, density of the slats (open area versus solid area), and the height of the bottom gap are all important in determining how efficiently the fence collects snow. By altering the height, density, and bottom gap, the designer can regulate to some degree, how the snow will accumulate behind the fence. If you have seen very many snow fences you will have noticed how they are often not aligned vertically. There is little loss of efficiency if they tilt in either direction from the vertical. In order to take greater advantage of the upstream decrease in wind speed produced by a single row of fences, a series of snow fences are often used

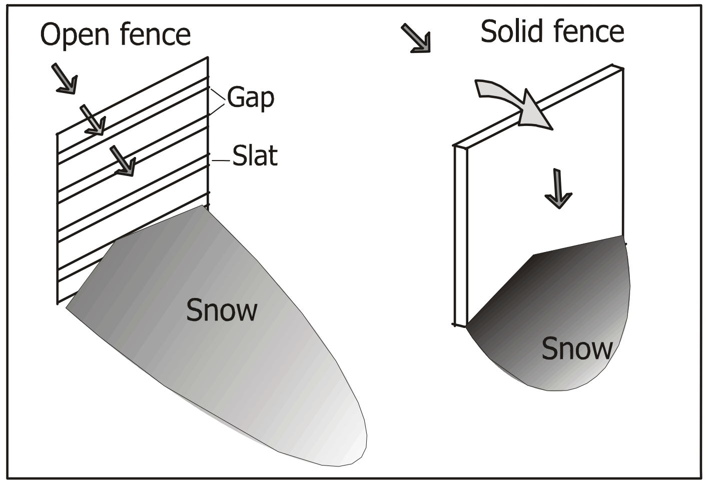
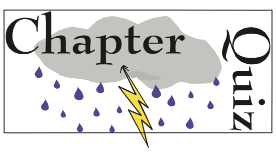


Figure 4.15- Two types of snow fences.

In terrain that has a high avalanche potential, fences are often seen high on the ridges. Strategic placement of fences is critical if they are going to be effective at decreasing the amount of snow that accumulates in the avalanche starting zone. The designer needs to know the direction(s) of the prevailing winds during snowstorms, and what slopes have the highest potential to slide.



1. Name the three jet streams?
2. True/False: Glacier winds only occur at night?
3. Barrier jets form when the lower atmosphere is ____________.
4. True/False: Downslope wind storms only form when the upstream wind is nearly parallel with the mountain range.
5. True/False: Valley winds develop faster than slope winds?
6. True/False: The stronger the bond between ice particles, the more difficult it is for the wind to transport it.
7. Slope winds take about ________ minutes to develop after sunlight 'heats' the air?
8. True/False: Downslope wind storms are common in the western Rockies?