Figure 1.8. Forest-cover damage by avalanche initiation in a logged clear-cut. (Photo by R. Laurilla)

Modern approaches to hazard mapping and zoning are considered best when based on risk concepts involving the three major components of risk: frequency of events, consequences of these events, and exposure of people and facilities in time and space. In this regard, Canada (CAA, 2002) has taken the lead by producing risk-based guidelines and standards for hazard mapping and zoning in terms of risk acceptability. One benefit is that risk-based guidelines for snow avalanches are compatible with risk-based guidelines for other mountain-slope hazards. All

about explosive control or avalanche-defense work at facilities sites. Cost-benefit analyses and value-engineering meetings are increasingly used to evaluate the options. Such studies provide needed guidance to avoid excessive costs and to eliminate inflexible options where flexibility is needed. When avalanche protection is planned for, economy and efficiency must be considered together, not just one or the other.

What lies in the future? More fatalities, more technological advances, and increasing risk could start a list of guesses, but such guesses are based on hindsight, which is often a poor predictor. Those enthralled with technology as the savior might do well to remember Pablo Picasso's opinion of computers: "For me, they are basically useless because all they can do is to provide solutions. What we need are people who can ask the right questions."

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Swiss Federal Institute for Snow and Avalanche

ELEMENTS OF MOUNTAIN SNOW CLIMATES AND WEATHER

The virtuous find delight in mountains,
the wise in rivers.
—Confucius

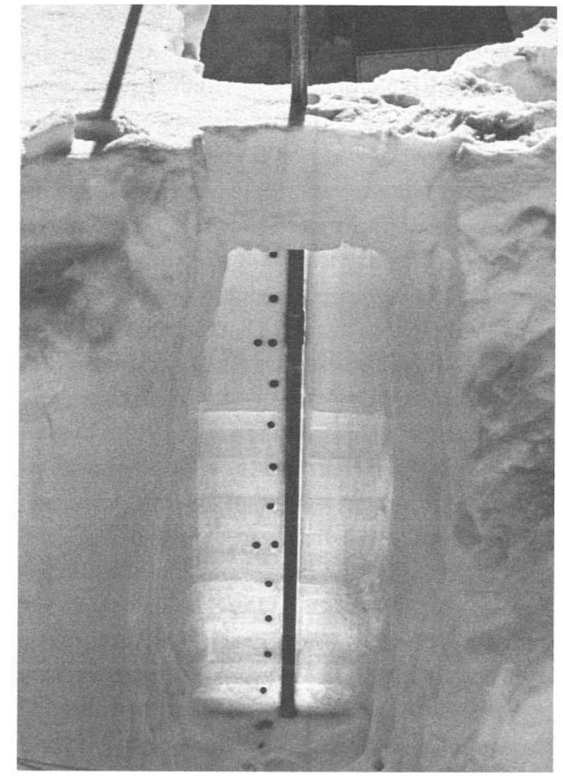

Figure 2.1. Illustration of snowpack structure. Structure usually depends on climate type in a general sense. (Swiss Federal Institute for Snow and Avalanche Research archive photo)

MOUNTAIN WEATHER AND SNOW-CLIMATE TYPES

Snow layering that contributes to avalanche formation is due to a combination of weather elements interacting with the snowpack. Most destructive avalanche cycles are caused by direct loading of snowfall from large-scale (synoptic) weather systems. To understand how, why, and when avalanches form, it is necessary to have some background about basic mountain weather, the effects of weather parameters on snow, and the interaction of weather and topography, which influence the deposition and distribution of snow. The primary weather and atmospheric factors contributing to avalanche formation include precipitation patterns and intensity, wind direction and wind speed, sensible heat, and radiation heating or cooling on snow.

Climate is defined as the average weather at a place. When avalanche formation is described in a broad, general sense, it is useful to classify the snow climate of a given mountain range as either maritime or continental (Figure 2.1). This classification will not be useful in the individual situations encountered in avalanche prediction. The reason is that avalanche prediction is dynamic and highly time-dependent (Chapter 6), whereas climate is only an average. Formation of dangerous layers requires a sequence of events that is in conflict with the average conditions embodied in terms of climate. The term "avalanche climate" has no meaning. However, in a *general* sense, the character of snow avalanching in a given range of mountains can be described as one of two basic

ing avalanche prediction in *general* terms. Problems such as land-use planning in snow-avalanche terrain (Chapters 5 and 10) may require a knowledge of snow climate since they involve time scales (or return periods of avalanches) on the order of decades to several hundred years. In the United States, it is popular to

Maritime Snow Climate

A maritime snow climate is characterized by relatively heavy snowfall and relatively mild temperatures. Snow covers are deep. While rain may fall at any time during the winter, cold, arctic air can also appear several times per winter. Maritime snow covers are often very unstable with rapidly fluctuating instability. Typical examples of maritime ranges include the Cascade Range in the western United States, the Coast Mountains of British Columbia, and the mountains of western Norway. The average annual snowfall in the maritime ranges of North America is 15 to 25 m.

Avalanche formation in maritime snow climates usually takes place during or immediately following storms, with failures occurring in the new snow near the surface. The prevalence of warm air temperatures promotes rapid stabilization of the snow near the surface once it falls, thereby limiting the time over which instability persists. A significant cause of major avalanching can be rain if it immediately follows deep, new snowfall. Rainfall may also cause formation of ice layers, which can act as future sliding layers when buried by subsequent snowstorms. Due to the deep snow covers and warm snowpack temperatures, the persistence of buried structural weaknesses deep in the snowpack is not usually as common in maritime snow climates as in continental snow climates. Weather observations are primary tools for predicting avalanches in a maritime snow climate (Table 2.1).

Continental Snow Climate

A continental snow climate is characterized by relatively low snowfall, cold temperatures, and a location considerably inland from coastal areas. Snow covers are relatively shallow and often unstable due to the persistence of structural weaknesses. Typical examples of continental ranges include the Rocky Mountains (Canadian and Colorado), the Brooks Range of Alaska, and the Pamirs of Asia. The annual snowfall in the continental ranges of North America is usually less than 8 m. As with any snow climate, most avalanches are caused by snow loading (new snow or drifting snow), and most form within new snow. Major avalanches, however, are often linked to buried structural weaknesses, which persist for long periods with, on average, a different character than in maritime snow climates.

Avalanche forecasting in continental ranges normally (i.e., during every winter) relies on observations of buried structural weaknesses in the snow cover as well as weather conditions that cause the failure of these layers. Many failures occur in new snow as well, but failure of old layers is a distinguishing feature of a continental snow climate. Extensive drifting in fair weather can cause heavy local accumulations to initiate failure in old snow layers. Due to the prevalence of thin snow covers, recrystallization weakens old snow in the presence of cold temperatures and high temperature gradients (see Chapter 3). The low temperatures also allow structural weaknesses to persist along with the basic anisotropy of the layers: weak in shear and resistant to settlement. Often in a continental range, it is possible to find nearly the entire snowpack consisting of weak faceted crystals (Chapter 3) and depth hoar grown under high temperature gradients, at least where the snowpack is thin.

The two broad classes are useful as a framework

Table 2.1 Characteristics of Maritime, Transitional, and Continental Snow Climates*

Type	Total Precipitation (mm)	Air Temperature (°C)	Snow Depth (cm)	New Snow Density (kg/m³)
Maritime	1,280	-1.3	190	120
Transitional	850	-4.7	170	90
Continental	550	-7.3	110	70

*Mean values compiled from 15 winters of U.S. data (Armstrong and Armstrong 1987).

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The two broad classes are useful as a framework for discussion of avalanche formation, but it must be stressed that considerable overlap occurs, with many areas exhibiting transitional features associated with both classes. That is, maritime conditions can occur in a continental range and vice versa. Examples of ranges that usually display transitional features (somewhere between maritime and continental) are the Selkirk Range of British Columbia, the Wasatch Range of Utah, and parts of the European Alps.

In any discussion of snow climate, altitude dependence has to be considered. Mountaintops in the high mountains will have different snow climates than lower down. Thus, one should think of snow climate and the character of avalanching associated with it only in broad, general terms on a synoptic-scale to characterize a mountain range. Local variations at smaller scales may show maritime character even though the main range is described as transitional. This spatial variation is another reason that snow climate and the associated character cannot be used to forecast avalanches.

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Type	Total Precipitation (mm)	Air Temperature (°C)	Snow Depth (cm)	New Snow Density (kg/m³)
Maritime	1,280	-1.3	190	120
[illegible]	[illegible]	[illegible]	170	90

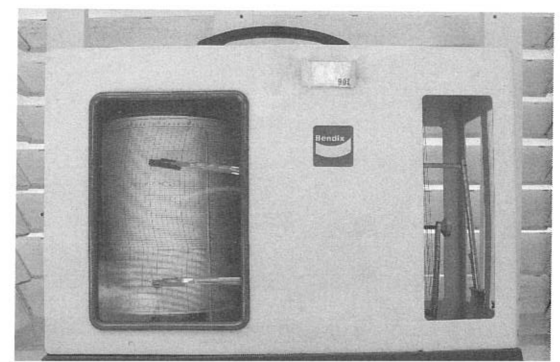
MOUNTAIN WIND AND PRECIPITATION

Wind speed and direction and the resulting precipitation patterns are crucial elements in forecasting avalanches. In general, wind direction and speed depend on the balance of forces on the moving air. The horizontal component of the wind velocity is the most important element in terms of wind speed and direction in mountainous avalanche terrain; the vertical component is the key factor in determining the amount and pattern of snowfall.

Horizontal Wind Component: Direction and Speed

The character of the horizontal component of the wind field depends strongly on altitude because the mountains act as huge, roughness obstacles to the wind blowing over them. The atmospheric pressure is about 1,000 millibars (mb) at sea level (1,013 mb is one standard atmosphere) (Figure 2.2a, b) and near the Earth's surface largely absent at about 1,000 m above the surface. The atmospheric pressure level of interest in avalanche forecasting depends on the height of the mountaintops (Figure 2.4). In the United States and Canada, the usual range of interest varies from 850 to 700 mb (excluding the highest ranges). This range is comparable to heights of 1,500 to 3,000 m. At elevations of avalanche-starting zones, frictional forces are still important in controlling the wind speed and direction in mountainous terrain. In North American ranges at the 700-mb level, for example, the forces controlling the wind speed and direction are (1) atmospheric-pressure changes (gradients) associated with general (synoptic) weather conditions; (2) the frictional force due to drag from the Earth's terrain features (directed opposite to the wind direction); and (3) an apparent force (Coriolis force) due to the Earth's rotation that is at right angles to the wind (to the right in the Northern Hemisphere, to the left in the Southern Hemisphere). In high-mountain terrain, the balance of these forces causes the wind to blow at an angle to the isobars (lines of constant pressure) toward lower pressure at a speed determined by the exact balance of forces.

Free-air wind speeds are those well above rough mountain topography (for example, several hundred meters or more) so that frictional forces are absent. Most mountain weather and avalanche-forecast operations predict either free-air wind speeds and directions



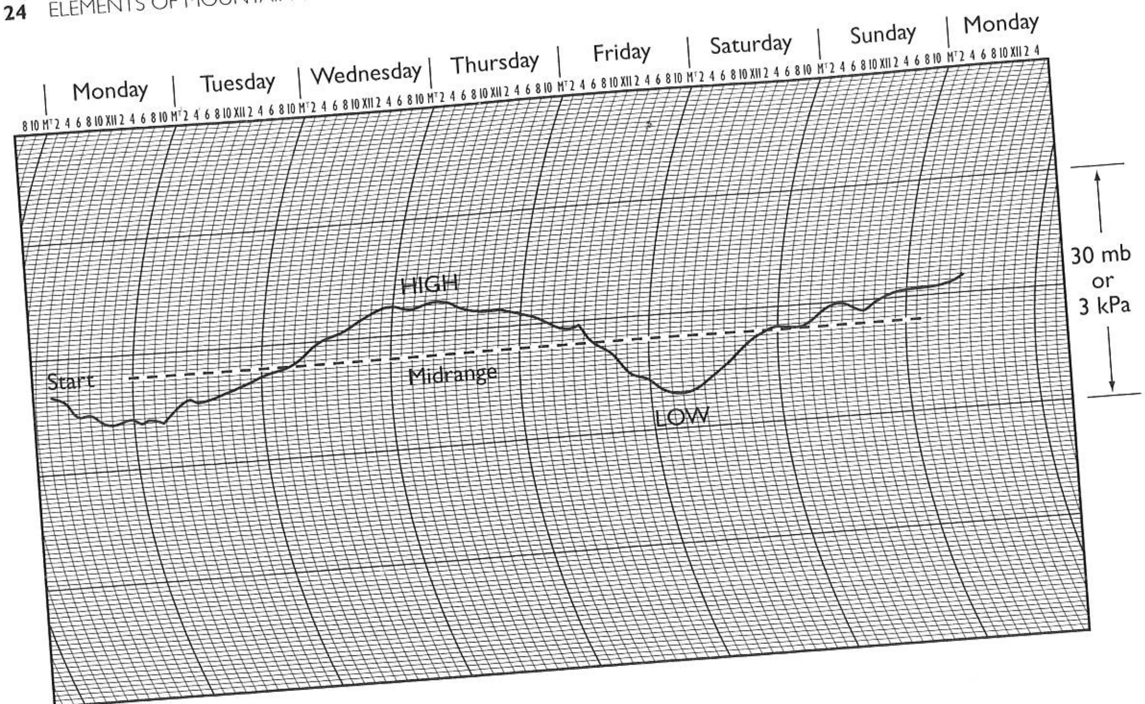


Figure 2.2b. Chart with pressure ranges identified.

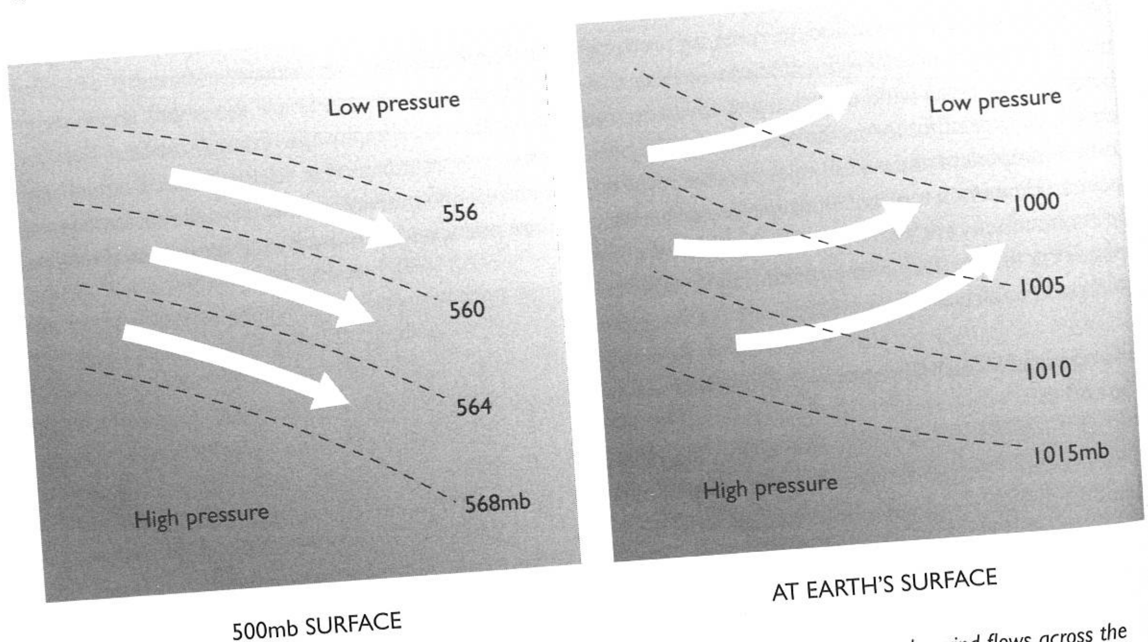


Figure 2.3. At 500mb the wind (free-air speed) is parallel to pressure contours. At the surface, the wind flows across the contours.

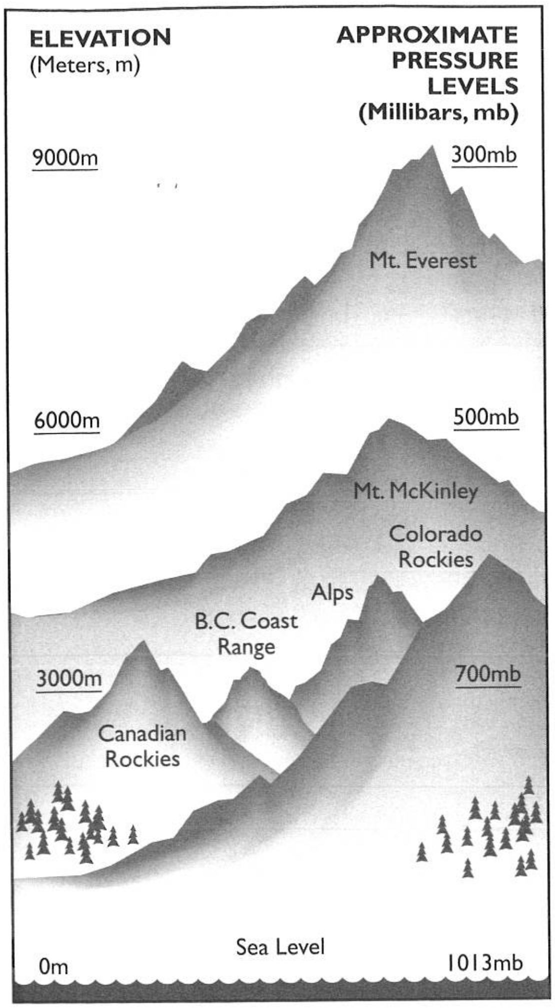


Figure 2.4. Atmospheric pressure variations with altitude.

(for high-elevation sites) or surface winds (for example, at lower elevations such as mountain passes). It is then up to the forecaster or field practitioner to determine the influences of the topography on these winds at the local scale.

Wind-speed sensors are placed on ridge tops, mountainsides, or in valleys. Therefore, they provide measurements influenced by frictional forces. Frictional forces have a strong influence on wind direction. In some cases, when a mountain range forms a barrier between warm maritime air and cold continental air, free-air winds at upper levels can blow at 180° to the direction of surface pressure–gradient winds, which have been strongly channeled through ground topography.

Vertical Wind Component: Amount, Rate, and Pattern of Precipitation

The vertical component of the wind causes precipitation; it determines how much, the rate, and the distribution. The precipitation rate is approximately proportional to the vertical component of the wind velocity. Therefore, the vertical wind speed is crucial for quantitative-snowfall forecasts, which are an essential element of avalanche forecasting.

In general, processes that cause an upward component of the air motion result in cloud formation and precipitation; those with a downward component dissipate clouds and produce clearing skies. When moisture-laden air moves up a mountain, its temperature decreases with height due to expansion as it moves through progressively decreasing pressure. In general, the amount of temperature decrease that an air parcel experiences with height depends on how much moisture is contained in it. Typical values are 1°C temperature change per 100 m of elevation (1°C/100 m) for dry air and about half that rate for moist air. Whether a vertically displaced air parcel is warmer or colder than the surrounding air will depend on the *average lapse rate* (that is, the temperature decrease with height in the surrounding air). The average lapse rate in the lowest 10 km of the atmosphere is about 6.5°C/1,000 m. Since cold air is denser than warm air, an air parcel will sink when it is colder than the surrounding air. When a parcel is warmer than the surrounding air, it will rise, expand, and cool during ascent. Eventually, the dew point will be reached, the air becomes moisture saturated, and further cooling will result in cloud formation through condensation of water vapor. The amount of moisture that air can hold *before* it condenses depends on its temperature: A parcel of warm air can potentially hold more water vapor prior to condensation than one with cold air. Further lifting causes water-droplet growth around condensation nuclei (see Chapter 3) and then precipitation. In summary, the rate at which air rises, the

Table 2.2 Characteristics of Major Lifting Types (Typical Values)

Type	Vertical Wind Speed	Precipitation Rate (water equivalent) (mm/h)	Duration of Precipitation	Horizontal Scale (km)
Cyclonic	~1–10 cm/s	Up to 2	Tens of hours to several days	1,000
Frontal	~1–20 cm/s	~1–10	Up to tens of hours	100 for width, 1,000 for length
Orographic	~1 cm/s–2 m/s	~1–5	Up to tens of hours	10–100
Convective	~1–10 m/s	~1–30	Minutes to hours	0.1–10

amount of moisture it contains, and its initial temperature are crucial determining factors for the formation and amount of precipitation.

In general, four mechanisms cause air to rise, and the vertical rate of ascent strongly depends on which mechanism is responsible. These mechanisms are (1) cyclonic convergence—upward air motion around a center of surface low pressure; (2) frontal lifting—upward motion, when one air mass is forced to rise over another; (3) orographic lifting—topographically induced, forced ascent; and (4) convection—thermally induced vertical motion. As a rough estimate, the following percentages apply to winter precipitation totals for the mechanisms: cyclonic, 10%; frontal, 30%; orographic, 50%; convection, 5%. These values will change with the season, with mountain climate, and from year to year. In most cases, more than one mechanism operates at a time, complicating precipitation forecasting. Table 2.2 gives typical characteristics of the mechanisms.

CONVERGENCE: UPWARD MOTION AROUND A LOW-PRESSURE AREA

Convergence around a low-pressure area (cyclone) will result in a general vertical motion of the air (Figure 2.5). In the Northern Hemisphere, the air circulates counterclockwise and inward toward the center of the storm (low pressure). To accommodate this inward flow, the air in the cyclone must then be forced up as the horizontal dimension of the air mass decreases (converges). The usual result is cloud formation over a large area and, in general, widespread precipitation over the mountains if they happen to be

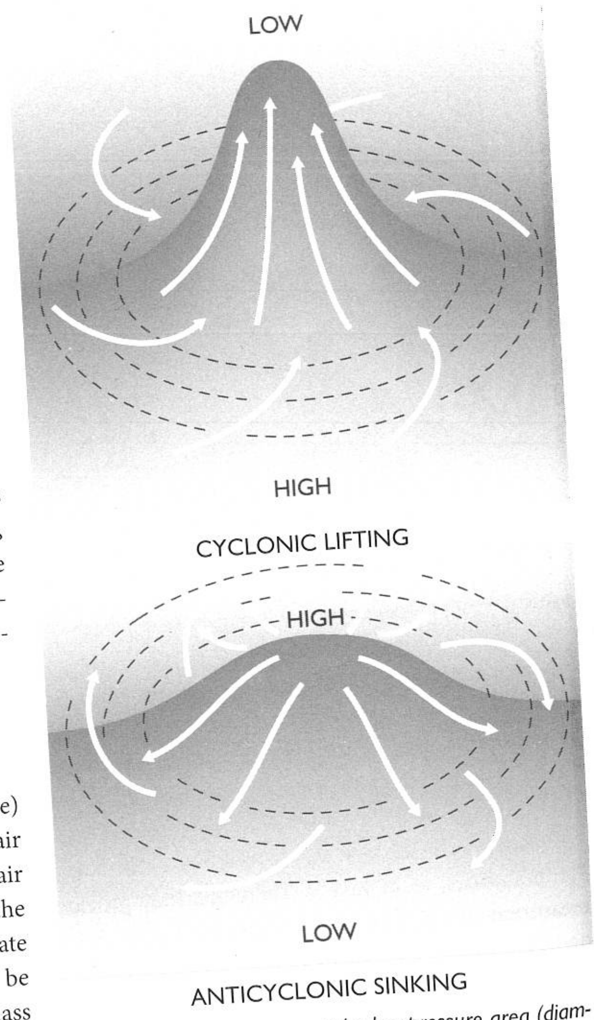


Figure 2.5. Convergence around a low-pressure area (diameter of about 1,000 km) causes widespread precipitation. Divergence (sinking) around a high causes clearing skies.

located near the low-pressure center. The opposite effect occurs in the case of divergence, resulting in sinking of the air downward and out from a region of high pressure, resulting in drying and cloud dissipation.

FRONTAL LIFTING

Another lifting mechanism is associated with the motion of an atmospheric front. A front is the boundary between air masses of different temperature (Figure 2.6). The boundary always slopes upward over the colder air. While the slope of the boundary is typically 1:100 or less for warm fronts, that for cold fronts is much steeper (as high as 1:25). The difference in frontal slopes leads to important variations in the rate, duration, and type of snow. Warm-front precipitation may last 12 to 18 hours or more. Cold-front precipitation typically occurs in a series of shower bands oriented almost parallel to the front with intense precipitation for about 4 to 6 hours.

When warm air moves against a wedge of cold air (warm-front passage) or if a wedge of cold air pushes under warm air (cold-front passage), forced lifting occurs, which causes cooling by expansion, leading to cloud formation and precipitation. In this case, the precipitation occurs over the frontal surface (typically 50 to 100 km wide). The occurrence is usually more widespread during warm-front passage and more sharply defined during cold-front passage. Mountain ranges can trap cold-air masses on their windward sides, or the pressure drop associated with the warm front may draw cold air over the mountains toward the front. Both of these mechanisms can intensify the lifting effect on warm fronts. The vertical wind-velocity component induced by frontal lifting is usually small and comparable to lifting rates from convergence.

OROGRAPHIC LIFTING

When moist air is forced over a mountain range by horizontal pressure differences, it is forced to flow parallel to the slope of the mountain (Figure 2.7). This causes an upward (vertical) component of the velocity, which is a significant fraction of the incoming horizontal wind speed. The vertical wind speed

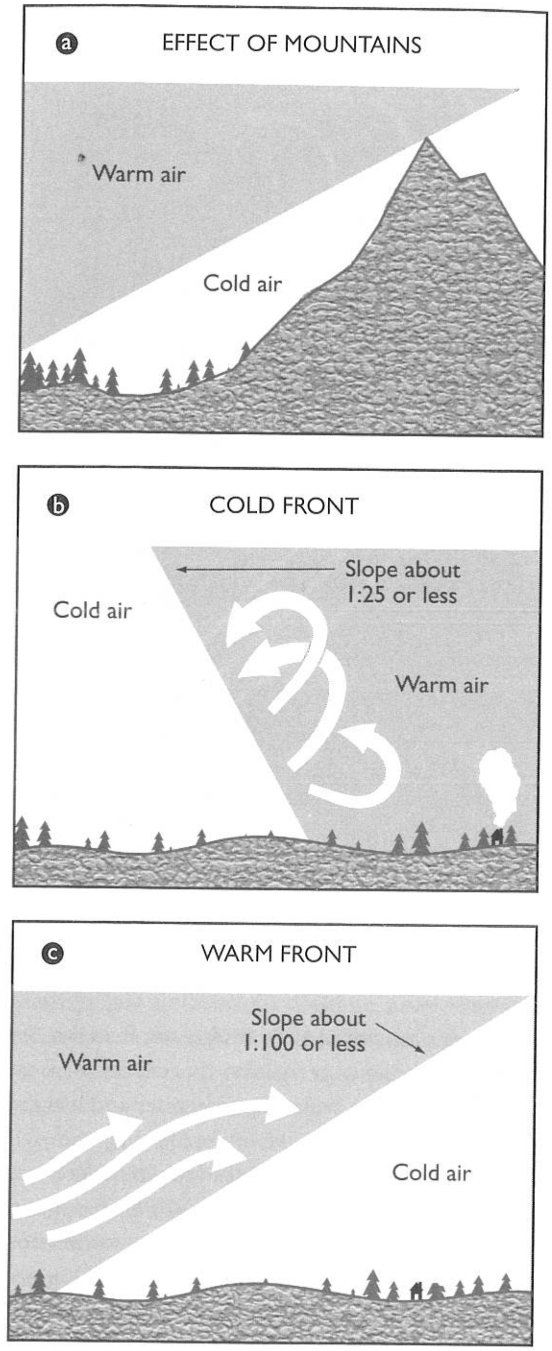


Figure 2.6. Frontal motion and the lifting effect of mountains on warm fronts. a and b: Due to differences in slope, precipitation is much more widespread during warm-front passage. c: Mountains increase the lifting of warm fronts.

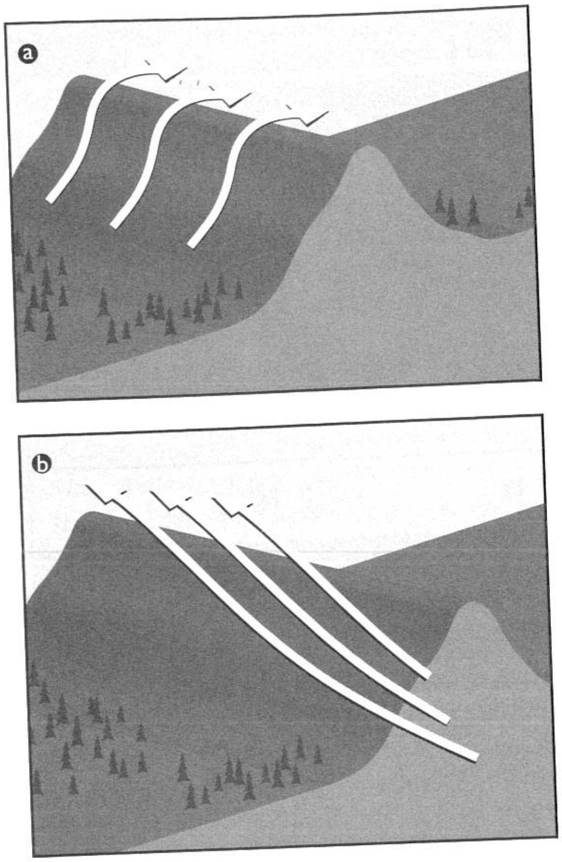


Figure 2.7. Orographic lifting is the most important winter precipitation mechanism; maximum effect is produced when the wind is perpendicular to the mountain barrier (a).

produced is a factor of 10 to 100 greater than that for frontal lifting and convergence. This effect is of great importance for the avalanche forecaster and it is the primary consideration when forecasting snowfall amounts in mountain areas. Orographic effects are estimated to account for 50% to 70% of mountain precipitation in winter. In general, the steeper the mountain slope that the wind strikes and the more directly perpendicular the wind hits the mountain, the greater the potential for producing precipitation. Maximum precipitation is achieved when the horizontal wind is blowing perpendicular to the ridge crest so the vertical component of the velocity produced is a maximum.

CONVECTION

Convection results when air is heated near the Earth's surface and is forced to rise because of its lower density relative to the surrounding air. Convection is a very local effect and its direct contribution to winter precipitation is small. However, convective instability following the passage of cold marine air over warm ocean water can contribute significantly to precipitation in coastal mountains. The resulting clouds and precipitation then contribute to the orographic component as the moisture is forced over the mountains. Convection can also be an important contributor to summer precipitation patterns in continental climates.

QUANTITATIVE PRECIPITATION FORECASTS

Avalanche and mountain weather forecasters operating out of weather offices produce quantitative-precipitation forecasts to aid avalanche prediction. Forecasters have access to a host of models, mostly implemented on computers, to determine vertical wind velocities due to fronts, convection, orographic effects, and general circulation. In addition, experience relevant to the mountain range in question is often integrated with other tools, such as satellite imagery, climatology records, remote weather-station output, radiosonde soundings, radar, and standard weather-station products. In most cases, several precipitation mechanisms operate simultaneously to complicate the prediction scheme. With the rapid increases in computer power, modern computing supplies weather forecasts with grid spacing on the order of 1 km, which is a high-enough resolution to include mountain-terrain features in the calcuations. This will change rapidly in the future as available computer power increases.

The prediction of snowfall amounts and rate are vital to avalanche forecasts. However, the prediction skills for precipitation, in general, have far more error associated with them than other variables, such as wind and air temperatures, which can be calculated from fluid models. For avalanche forecasting, precipitation must be quantified on the scale of a mountain-range width or less (meso-scale, 1 to 100 km), whereas precipitation amounts are normally forecast on a synoptic scale (storm width, 1,000 km) by conventional weather forecasters. Therefore, avalanche weather forecasters use *orographic precipitation models* to predict precipitation amounts as storms force moisture-laden air through the mountains.

OROGRAPHIC PRECIPITATION MODELS

Orographic precipitation models include the assumption that precipitation is produced at a rate that is directly proportional to the rate at which the air is lifted (vertical component of wind velocity) over the mountains. The first mountain struck will usually induce the most precipitation and subsequent barriers receive less as the moisture supply in the air mass diminishes.

Orographic models include computer-generated maps of mountain terrain necessary to predict areas of lifting, subsidence, and shadowing (e.g., 155,000 km^2 in Colorado). The procedure involves keeping track of the amount of condensation and precipitation as an air mass rises over the mountains. On the downwind side of the crest, much of the condensed moisture evaporates.

Orographic precipitation models require initial data from the forecaster before they are run. These data include wind direction, wind speed, a temperature profile (lapse rate), and an estimate of the thickness and width of the atmospheric layer capable of producing precipitation. These data are obtained from upper air soundings and standard weather-forecast products. The duration of the precipitation is obtained by estimating the width of the moist layer and wind speed.

In addition to lifting mechanisms, other (local) effects are important in mountain-precipitation forecasting:

- *Topographic convergence* refers to the flowing together and subsequent lifting of moist air downwind of a major mountain barrier. One well-known example is the Puget Sound convergence zone in Washington State. This phenomenon results from splitting and reconverging of the airflow around the Olympic Mountains west of Seattle. Subsequent forced lifting of the convergence-produced clouds and moisture results in enhanced snowfall in the Cascade Range east of Seattle. Topographic convergence can result in extremely variable precipitation patterns. Variations in snowfall amounts by a factor of 10 over a distance of less than 1 km have been observed. Local topographic convergence effects have also been noted behind the volcanoes of the Cascade Range.
- *Orographic convergence* refers to local convergence zones that can shift or form and disappear as the prevailing wind direction changes with respect to a given topographic feature.
- *Valley channeling* is the forced convergence of air by a gradually narrowing valley. The result is enhanced precipitation in the narrow portions of the valley over that expected for constant valley width.

All three of these effects can have major influences on the local distribution of snow. Local knowledge of all three is essential for good mountain-precipitation forecasting. While an experienced forecaster may identify and partially quantify the expected precipitation patterns, success is not guaranteed.

In addition to orographic effects, other influences may cause snowfall to *decrease* with elevation, which can complicate the picture. Sometimes in coastal areas, an air mass can move slowly off the ocean with very little lifting to produce a band of precipitation at relatively low elevation. In addition, some observers cite evidence for enhanced snowfall near the freezing level, which may produce snowfall decrease with height.

LOCAL WIND FLOW OVER MOUNTAIN TERRAIN

Wind speed and direction in the mountains are crucial factors for determining whether or not avalanches will occur and the location and the character of avalanches that do form. In modern avalanche forecasting, wind speed and direction are considered essential input. On a local scale, wind effects are crucial for determining route selection in backcountry travel (Figure 2.8).

On an individual mountain, the characteristics

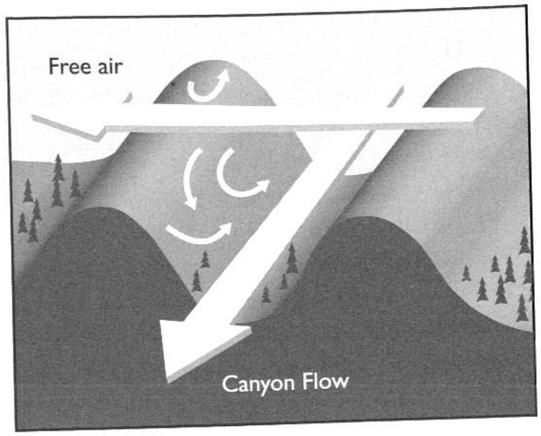


Figure 2.8. Deflection of flow in the direction of a canyon.

of wind velocity change with both altitude and the local topographic features. Generally, in the middle and high latitudes, wind speed increases with height due to the characteristics of the global westerly wind belts. The amount of snowfall generally *increases* with height on mountains of moderate height due to orographic effects (Figure 2.9). On the world's highest peaks, however, there is a limit beyond which precipitation decreases with height because previous lifting has caused most of the available precipitation to fall out. In addition, near the tops of exposed peaks or ridges, the winds are often so strong that snow is blown off. This occurs because of general increases in wind speed with elevation and also because topographical frictional effects are greatly reduced.

On a given mountain, snowfall is generally greater on the windward side (lifting occurs) than on the leeward side (subsidence occurs). In addition, as a storm passes over the mountain, snowfall is gradually subtracted from the total moisture available so that by the time the lee side is reached there is less moisture left to deposit. This same concept explains partly why continental ranges have less snowfall: Often, a lot of snow falls in coastal ranges before the precipitating storm reaches inland destinations.

The *local* terrain features on a mountain have very important effects on snow deposition and wind patterns. A key concept for avalanche formation is that, in general, snow is picked up in places where the wind accelerates, and snow is deposited in deceleration zones (Figure 2.10). The vertical compression of airflow over a mountain causes acceleration. Deceleration is caused by zones of low pressure (high turbulent intensity) and dynamic wakes at the snow surface. All of these effects can occur simultaneously, but on the windward side of a ridge crest, the wind flow is compressed (air pressure is higher) against the mountain, and acceleration (compressional effects) dominates. On the lee side of a mountain, the compressional effects are largely absent and deceleration dominates. Typically, a pressure difference of 1 mb results in an

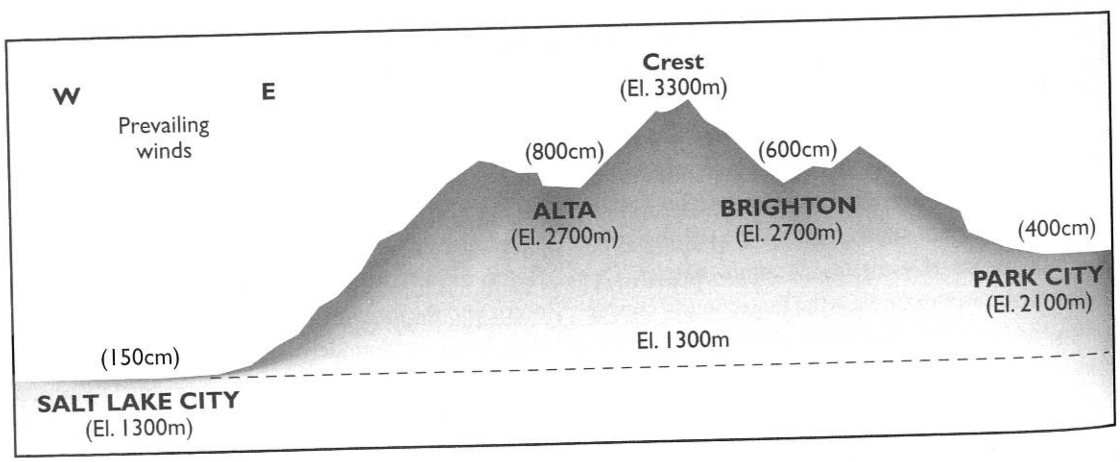


Figure 2.9. Snowfall varies with altitude and exposure to the prevailing wind within a mountain range.

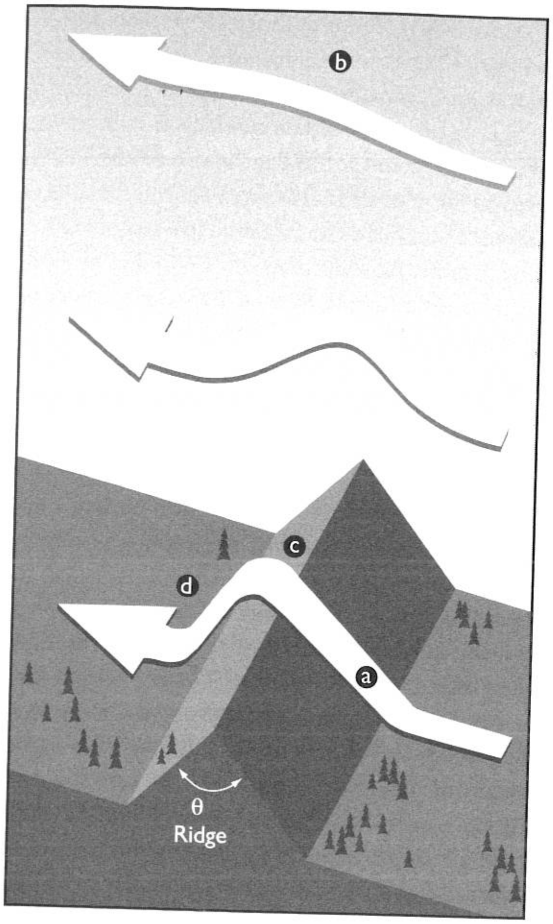


Figure 2.10. a: Airflow over a ridge, b: free-air motion, c: ridge crest, d: acceleration and deceleration zones.

increase in wind speed of 4 to 5 m/s over a ridge crest. It is common for the compression effect to cause winds on an exposed ridge crest to exceed the free-air speed. Sharp breaks of slope cause more turbulence in the air passing over them than if the slope change is gradual, due to greater pressure changes. This can cause the airflow to separate from the ground on the lee side; vertical eddies form and a reversal of flow direction at the snow surface takes place. These eddies are an important element in cornice formation, snow "dust" devils, and the determination of the distribution and character of deposited snow for avalanche formation (Figure 2.11).

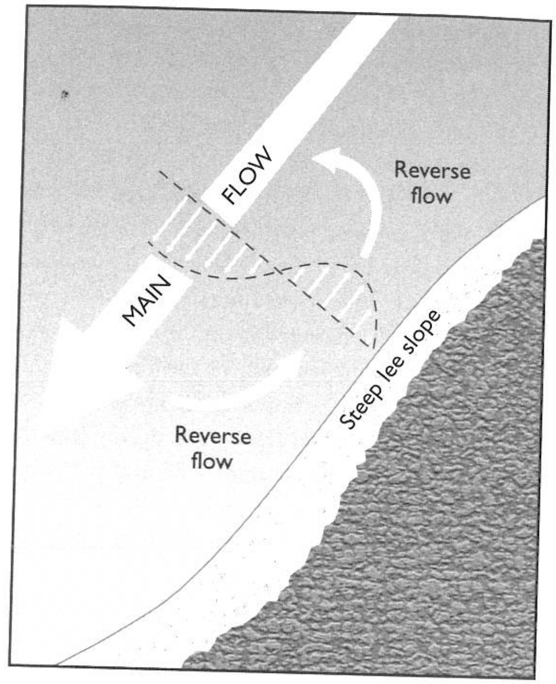


Figure 2.11. Reversal of flow on steep lee slopes. This pattern allows cornice and avalanche formation.

The same logic applies to cross-loaded slopes where the slope breaks are not sharp enough to allow cornice formation. Wind deposition of snow will still be on the lee side of a ridge, but the prevailing wind direction is not as easy for an observer to determine from a distance compared to when a cornice forms. Field observations and measurements show that minor rolls or changes in slope angle may significantly affect the character and areal distribution of wind-deposited snow.

Intense mountain winds are sometimes encountered down mountain canyons oriented perpendicular to the free air flowing across a mountain range. This effect can produce cross-slope loading by drifting snow (see Figure 2.8).

When a large-scale low-pressure area is present in the lee of a mountain range, sometimes a dry, warm *foehn wind* (or *chinook)* will blow down the leeward side of the mountain. The foehn wind occurs when the prevailing winds in warm, moist air are directed upward against the mountains by the large-scale pressure differences between the windward and leeward sides (Figure 2.12). The ascent usually causes cloud buildup and sometimes precipitation on the windward side. The precipitation on the windward side can, of course, add load to the snowpack. However, the main feature of a foehn wind is the intense flow of air, which dries and warms due to compression as it descends the leeward side. The intense warming and drying is largely a result of loss of precipitation on the windward side and descent at the dry lapse rate on the leeward side.

Foehn winds can be important in avalanche work due to the warming that takes place during leeward descent. They are an important source of wet snow avalanching in the European and New Zealand Alps. In general, any snow-temperature rise can adversely affect snow stability (see Chapter 7). However, since most avalanche-starting areas are near the tops of peaks and since the warming takes place due to descent, foehn winds will have little effect on warming at the highest starting areas. Refreezing of the snow surface after melt from a foehn can cause crust formation, which might serve as a sliding surface if buried by later snowfalls.

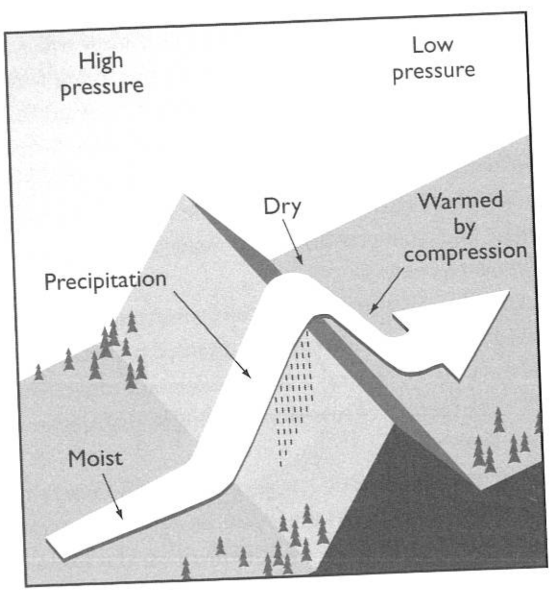


Figure 2.12. Schematic of foehn wind.

BLOWING AND DRIFTING SNOW

Redistribution of snow by wind is a major feature of mountain snowpacks and it is essential for avalanche formation in some instances. In fact, redistributed snow can account for avalanche releases by loading even in clear weather. The term *blowing snow* is reserved to describe particles raised to a height of about 2 m or more. Blowing snow often obscures visibility. Drifting snow (about 90% of transported snow) is used to describe near-surface transport.

The critical wind speed (threshold wind speed) at which snow is picked up from the surface by turbulent eddies of wind is a complicated function of the physical conditions of the surface snow. No guidelines have yet been published to quantify the conditions in a manner useful for avalanche workers. However, some useful general statements based on experimental information on the cohesion of ice and observation of blowing and drifting snow are possible: (1) Threshold wind speed increases with increasing temperature and humidity. (2) If the original deposition occurs with wind, the particles will be broken into small pieces and they will pack to a higher density to subsequently increase threshold wind speed. (3) Threshold wind speed will increase with time since deposition (due to bond formation between surface grains). The increase will slow with time and it is slower at colder temperatures (see Chapter 3 for ice-bond physics). (4) Threshold wind speed will be much lower if there is a source of particles, such as new snowfall, a low strength layer at the surface (e.g., surface hoar), or snow on trees.

For loose, unbonded snow, the typical threshold wind speed (at a 10-m height) is 5 m/s. For a dense, bonded snow cover, winds greater than 25 m/s are necessary to produce blowing snow. Blowing snow will occur with modest winds whenever snow is falling.

There are three modes of transport for wind-redistributed snow (Figure 2.13). *Rolling* involves the creeplike motion of dry particles along the surface (1-mm depth). *Saltation* occurs as particles bounce along the surface in a layer about 10 cm deep, dislodging other particles as they hit the surface. Saltation is

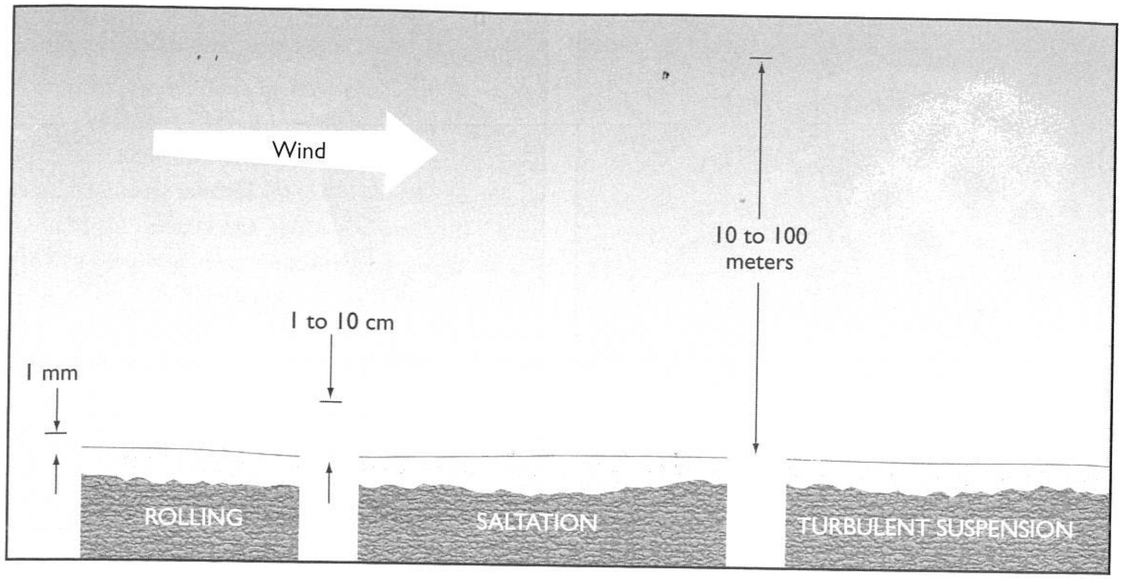


Figure 2.13. Modes of transport in blowing and drifting snow. (Modified from Mellor 1965)

initiated with winds of 5 to 10 m/s over cold, loose snow. Rolling is thought to account for about 10% of the mass when creep and saltation occur together. *Suspension* is caused by turbulent eddies lifting particles up to tens of meters above the surface. The transition from saltation to suspension occurs when the wind speed exceeds about 15 m/s. When turbulent upcurrents exceed the fall velocity of the snow particles (normally 0.2 to 2 m/s for newly fallen snow), experiments show that the majority of the mass is transported in the lowest meter above the surface. In general, turbulent suspension and saltation will both work to load avalanche slopes in regions of decelerating wind. With higher wind speed, saltation trajectories of particles will carry the particles higher, and the proportion of the load carried by suspension will increase.

In the mountains, snow redistribution is uneven because it is strongly influenced by the local topography, including vegetation, rock outcrops, the "caps" of gullies, cols, notches, and gully walls (Figures 2.14 and 2.15). Hollows 10 to 100 m across tend to be filled in during the course of a winter until an equilibrium level of the snow surface is reached, where erosion balances deposition. Therefore, the amount of material blown past hollows increases during the winter as hollows are filled in and the snow-surface topography changes.

Gullies are primary places for avalanche formation in the windy climate of Scotland because they are snow collectors and favored places for winter climbing. Snow instability can go from nonexistent

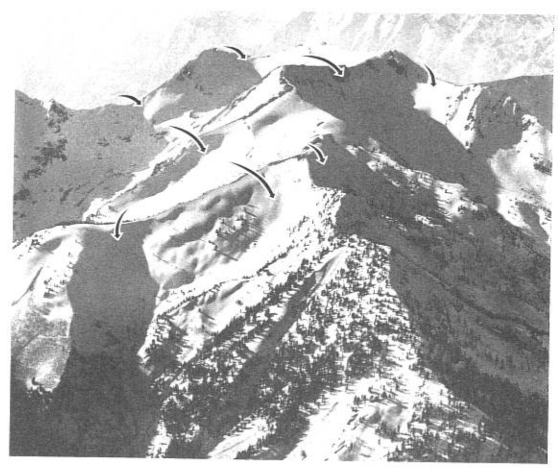


Figure 2.14. Likely locations for cross-loading, gully, and notch deposition (shown by arrows) depend on prevailing wind direction.

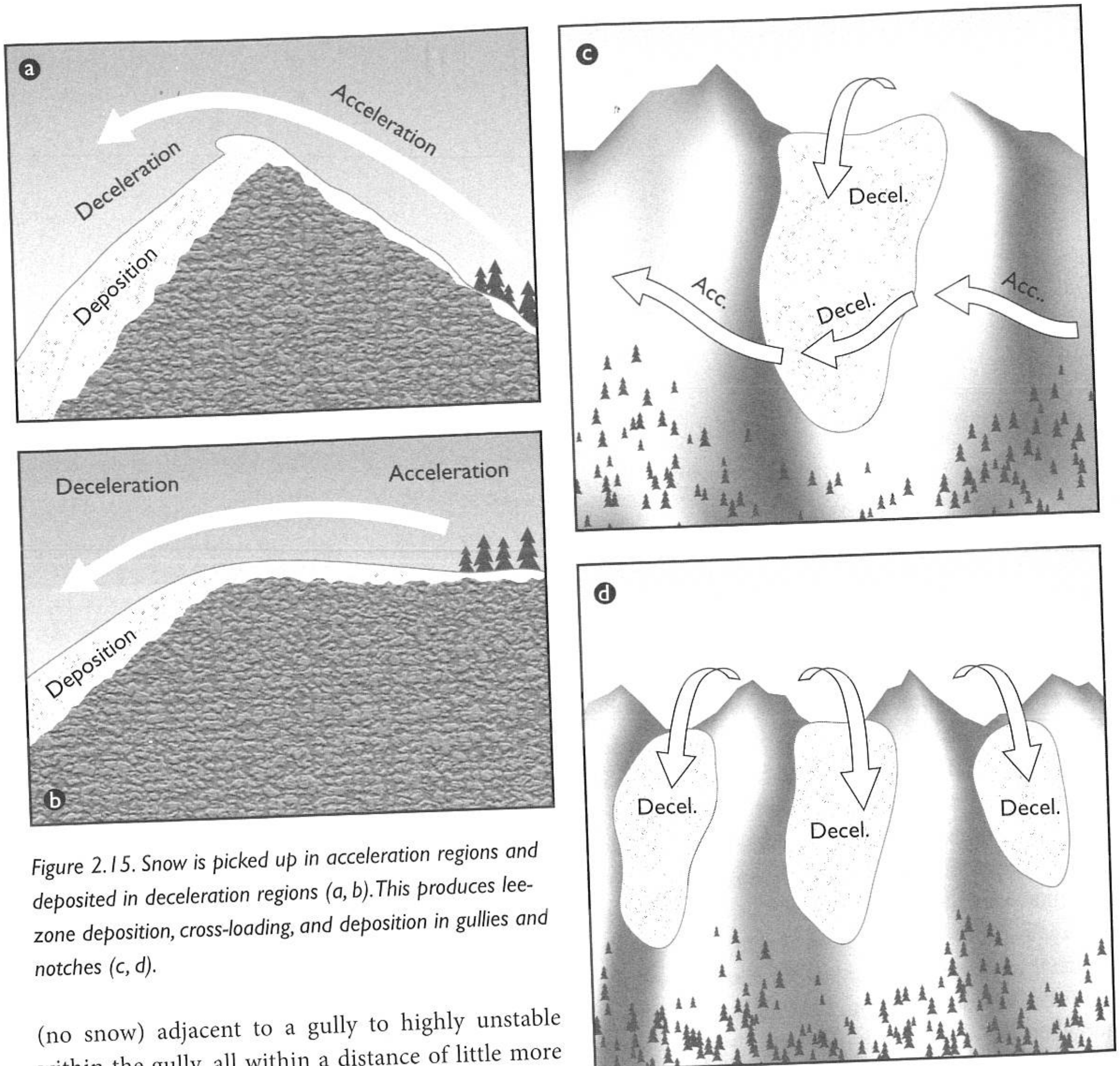


Figure 2.15. Snow is picked up in acceleration regions and deposited in deceleration regions (a, b). This produces lee-zone deposition, cross-loading, and deposition in gullies and notches (c, d).

(no snow) adjacent to a gully to highly unstable within the gully, all within a distance of little more than 1 m. Small-scale terrain features near crests of ridges where *both* the snow cover and terrain are changing rapidly have been highlighted as prominent in avalanches triggered by skiers (see Chapter 8). Combinations of small-scale terrain features, such as convex rolls near ridge crests, are shown by statistics of skier-triggered avalanches to be important sources of accidents.

Avalanche incidents (including at least one fatality) have been attributed to avalanche formation in "blow holes" sculpted behind large-terrain features. The minimum change-of-slope angle necessary to cause significant changes in drift development is about 10°. The choice of a safe route in mountainous terrain is often dependent on recognizing micro-scale features and their effects on drift development.

LEE-SLOPE DEPOSITION: AVALANCHE AND CORNICE FORMATION

On the lee side of alpine ridge crests, where a sharp change in slope angle occurs, cornices and avalanche deposits may form due to formation of eddies by flow separation. The windward slope angle is thought to be critical in determining whether a cornice or snowdrift will form, but clearly the overall change in slope angle is important. In addition, as snow is redistributed, the particles become broken and abraded as they impact the snow surface. (Natural concentrations in blowing snow are thought to be generally too small for particle interactions above the surface.) These small fragments, upon deposition, become tightly packed and rapidly produce a slablike texture as they bond to their neighbors. This mechanism is therefore extremely important to avalanche formation (see Chapter 4). In addition to being favored places for producing slablike texture, lee zones usually collect greater amounts of snow than nearby wind-protected valleys (factors of three to five times more and higher are not uncommon).

Cornices usually form on ridge crests but they can form at any place where a sharp change in slope angle is found (Figures 2.16 and 2.17). The threshold wind speed for cornice formation and growth is about the same as the threshold wind speed for transport over loose, cold snow (5 to 10 m/s). For winds in excess of 25 m/s, studies have shown that cornices can decrease in size due to windward scouring of the root. Because the snow in cornices is usually hard and dense (up to 500 kg/m^3), the threshold wind speed for scouring may be greater than 25 m/s. Clearly, variations may result due to surface conditions, temperature, humidity, and other factors, but it appears that cornice growth occurs when wind speeds are in the range of 5 to 25 m/s. Also, for wind speeds in excess of 25 m/s, turbulent suspension is expected to be the mechanism for snow transport, so that snow particles may be transported far

Figure 2.16. Blowing snow during cornice formation and lee-zone deposition. (Photo by D. Fesler)

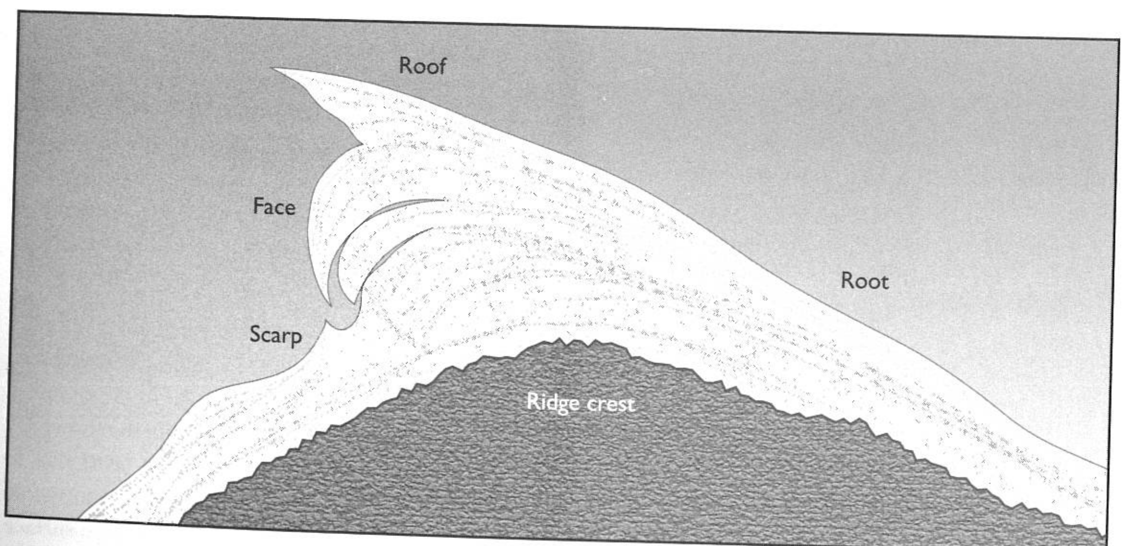


Figure 2.17. Cornice structure and nomenclature. Prevailing wind direction is from right to left.

beyond the cornice scarp and perhaps beyond potential avalanche-starting zones. In addition, such suspension can result in measurable amounts of snow being lost by sublimation as snow is transported through the air.

Cornices have three other features that are important to avalanche workers: (1) an overhanging cornice provides a quick assessment of the prevailing wind direction in a mountain range from a distance; (2) the steep, lee area below the face of a cornice is itself a prime area for unstable snow slabs to form (Figure 2.18); and (3) the overhanging face of a cornice on a ridge crest can and often does collapse. Many serious and fatal accidents have resulted from cornice collapses. Cornice collapse and sometimes avalanches triggered below them can often happen under daytime heating in good weather or, less commonly, from drifting snow during fair weather. Avalanches are triggered by cornices, and occasionally falling cornices have damaged structures. Ridge crests are preferred as travel routes when unstable snow is expected, but the overhanging roofs of cornices are an exception (Figure 2.19). The steeper the face below the cornice and the greater the overhang, the more unstable the cornice. At times, cornices may form on both sides of a ridge, causing further difficulty in travel.

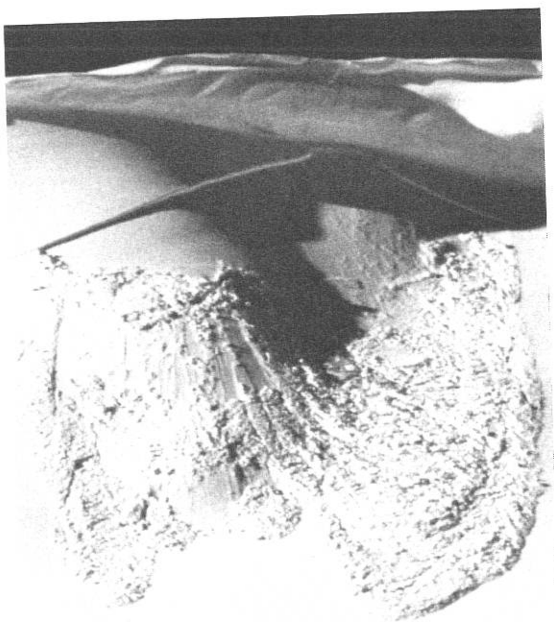


Figure 2.18. Slab avalanche formed under the lee of a cornice. (Photo by B. Jamieson)

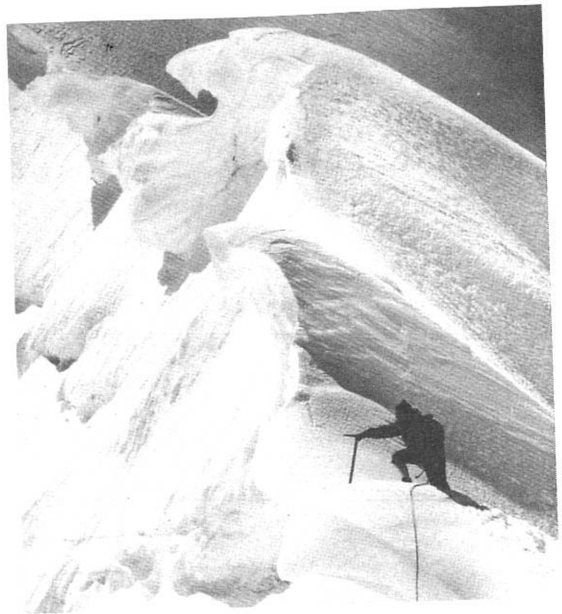


Figure 2.19. Cornice on a ridge crest. (Photo by A. Roch)

HEAT EXCHANGE AT THE SNOW SURFACE

The exchange of heat between the snow surface and the atmosphere is important for avalanche formation for both wet and dry snow. Heat exchange can alter surface snow to produce weak snow there, which may fail or act as a future failure layer when subsequently buried.

Heat can enter or leave the snowpack surface by conduction, convection, or radiation. Heat flow by conduction in the air is negligible with respect to other mechanisms because of the inefficiency of passing heat by molecular collisions in airflow (the thermal conductivity of air is extremely low).

Heat may be transferred to and from the snowpack by turbulent exchange (called *sensible heat*) due to wind eddies. If the air is warmer than the snowpack, surface heat is added to the snowpack. If the surface is warmer than the air, heat is lost from the snowpack.