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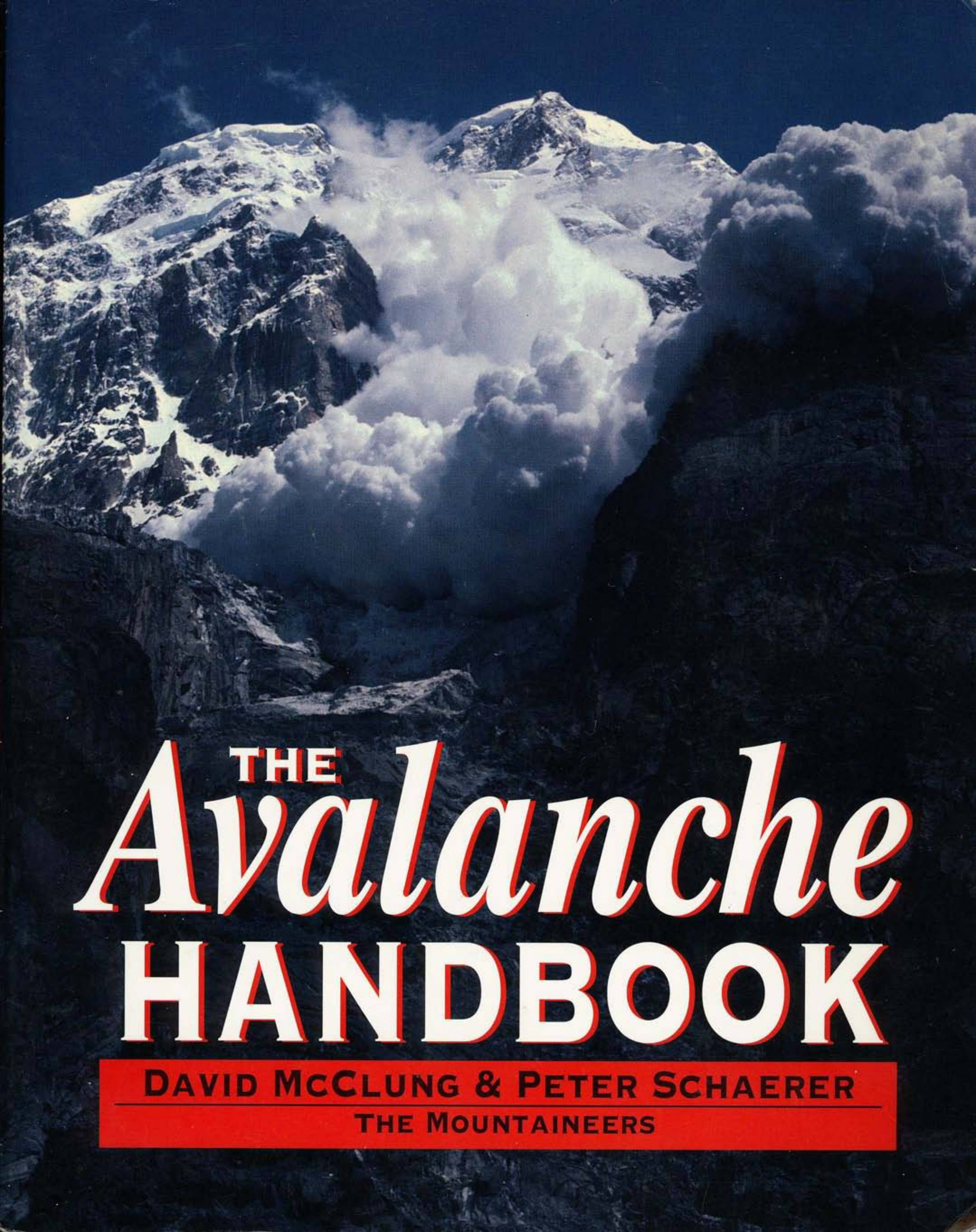

THE
Avalanche
HANDBOOK
DAVID McCLUNG & PETER SCHAERER
THE MOUNTAINEERS

CHAPTER 3
SNOW FORMATION AND GROWTH IN THE ATMOSPHERE AND SNOWPACK

You boil it in sawdust; you salt it in glue;
You condense it with locusts and tape;
Still keeping one principal object in view,
To preserve its symmetrical shape.

—Lewis Carroll

SNOW CRYSTAL FORMATION AND GROWTH IN THE ATMOSPHERE

Most avalanches occur as a result of newly fallen snow, perhaps more than 90% in some climates. Sometimes variations in type of new snow falling are responsible and experienced avalanche workers are alert to such changes.

Snow crystals begin their lives in atmospheric clouds. Clouds are composed of water droplets that form when the air is supersaturated with water vapor. The droplets form by condensation on small particles called *condensation nuclei* (salt, dust, or soil). These particles are very small with a typical diameter of 10^{-6} mm (1 μm), and they are always in abundant supply. Growth is by condensation of water vapor on their surfaces when the air is saturated with respect to the droplet (Figure 3.1).

When the air temperature at which the cloud becomes saturated is below 0°C, it is possible to form snow from tiny ice crystals. At these temperatures, small water droplets will remain as water droplets in a supercooled state. Typically these droplets are about 20 μm in size with concentrations of several hundred per cubic centimeter.

To form a small ice crystal by freezing, foreign particles are also needed around which the ice crystallizes. However, these ice crystal nuclei (*freezing nuclei*) are much less common than the condensation nuclei needed for forming water droplets. The typical size of freezing nuclei is the same as for condensation nuclei, but they have a special character that promotes freezing. Not all small particles (including dust, soil, and other chemical particles) are suitable for freezing nuclei; they must have the correct molecular structure. Also, freezing nuclei are variable with respect to the temperature at which they allow freezing to take place. The number of "active" freezing nuclei increases as the air temperature decreases. At –10°C there are about 10 active nuclei per cubic centimeter. As the temperature in the cloud decreases, it becomes much easier for ice crystals to form by freezing and the number of ice crystals increases relative to the number of droplets. At a temperature of –40°C, droplets will freeze by themselves without the aid of freezing nuclei.

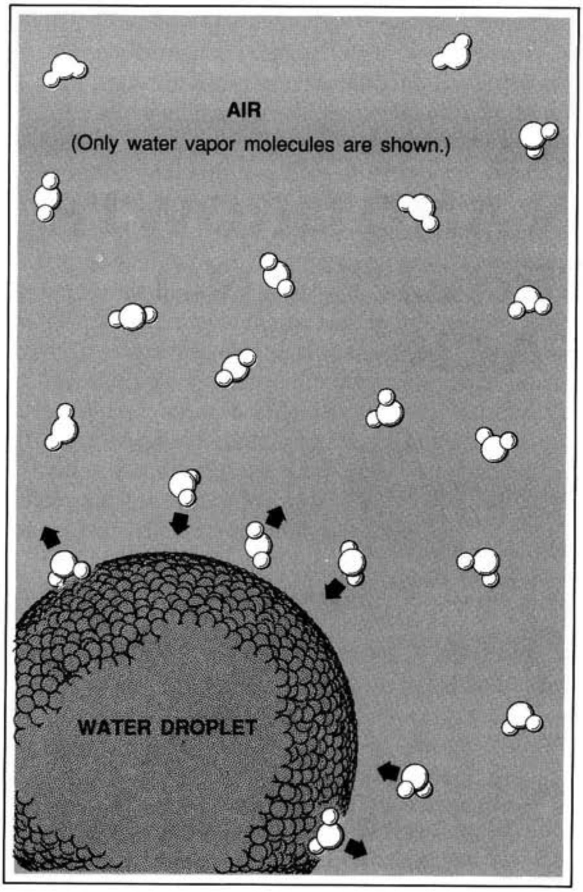


Figure 3.1. Growth of water droplets by condensation of water molecules when the air is supersaturated with respect to the droplet surface.

Once a small ice crystal forms, its subsequent growth is determined by two processes (Figure 3.2). The process that determines the basic crystal form occurs by direct transfer of water vapor molecules from the supercooled water droplets in the cloud. It has been determined experimentally and theoretically that the vapor pressure over a water droplet is higher than over an ice crystal at a given temperature (Figure 3.3). Since the

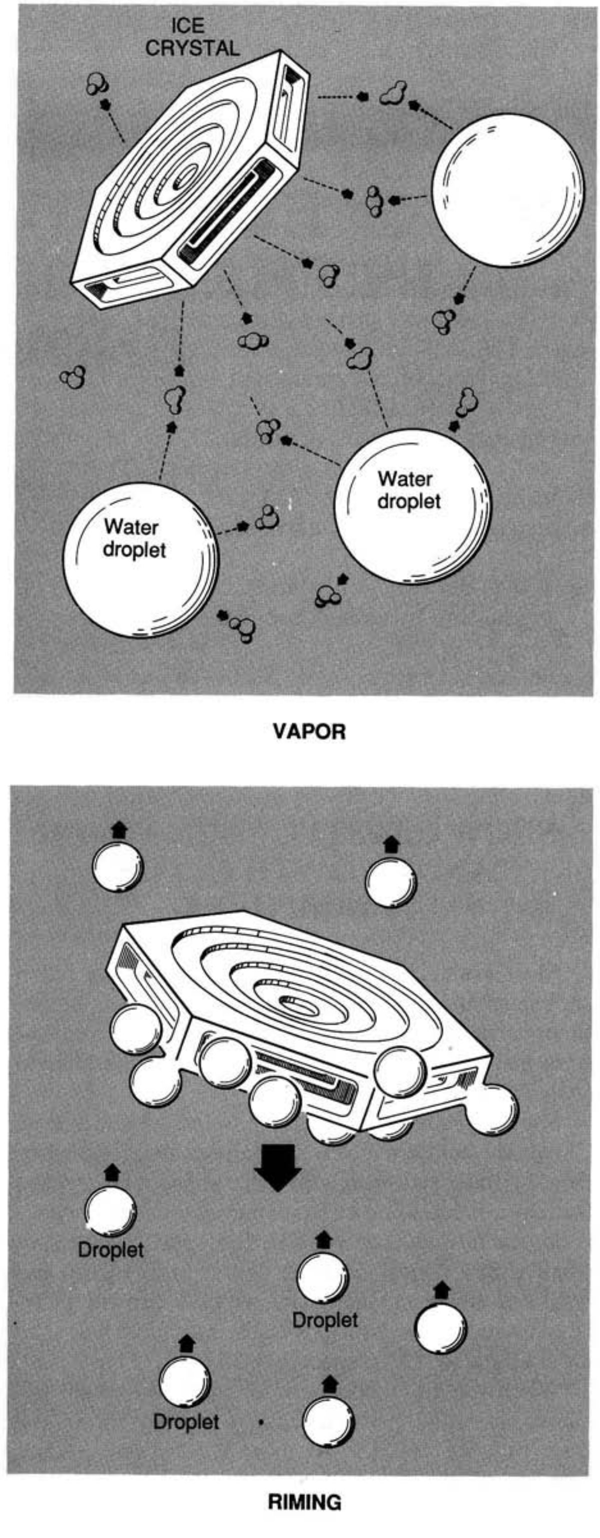


Figure 3.2. Two mechanisms of ice crystal growth in the atmosphere: transfer of molecules from droplets (top) and riming resulting from collisions with droplets during their fall through the atmosphere (bottom).

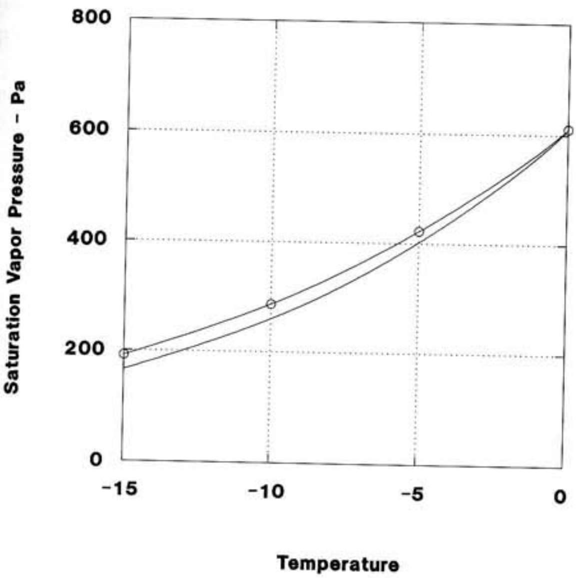


Figure 3.3. Saturation vapor pressure as a function of temperature over flat ice (bottom line) and water (top line) surfaces.

pressure is higher over the droplet, water vapor molecules diffuse toward neighboring ice crystals and they condense (deposit) from the vapor onto the ice crystal. Thus, ice crystals grow at the expense of supercooled droplets, due to vapor pressure differences between the droplets and the ice crystals.

The second growth mechanism occurs as the crystals move in the atmosphere. When the ice crystals attain a large enough size, they fall and gain mass by colliding with some of the larger supercooled droplets, which subsequently freeze onto the crystals in a second process called *riming*. The same process causes icing on an airplane wing moving through supercooled droplets in the atmosphere. When a crystal is rimed it usually falls faster because weight is added without much change in air resistance. In contrast, growth from the vapor usually increases air resistance by branch growth. Sometimes the crystal branches become entirely filled in by riming to form a rounded crystal called *graupel* in which the original type of crystal is usually unrecognizable (see Table C.1 in Appendix C). This requires a long growth period either from passage through thick clouds or repeated rides up and down in thermal convection updrafts in clouds that prolong the riming process. Graupel particles can also form hail if they ride updrafts involving freeze-thaw cycles.

The ultimate form that a snow crystal attains in the atmosphere depends on complicated conditions at and near the crystal surface, but temperature is the most important variable. In general, growth usually occurs in two basic directions: in the basal plane of the ice crystal or perpendicular to it. Ice crystals have three intrinsic axes in the basal plane (*a*-axes) separated by 120°, and an axis perpendicular to the basal plane (the *c*-axis) (see Figure 3.4). There is hexagonal symmetry in the basal plane and heat flows less efficiently in the basal plane

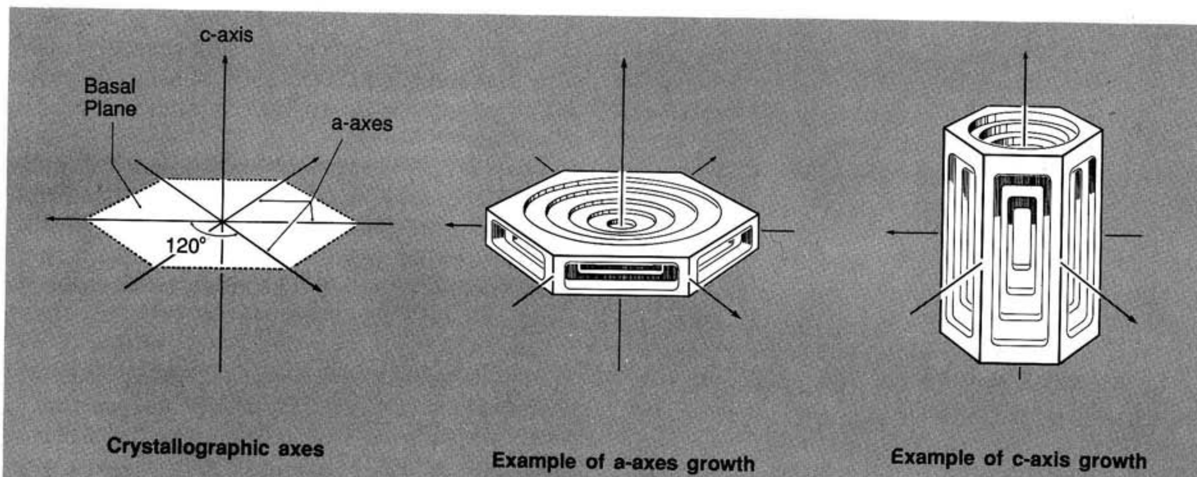


Figure 3.4. Crystallographic axes and examples of a*-axis and* c*-axis growth for ice crystals.*

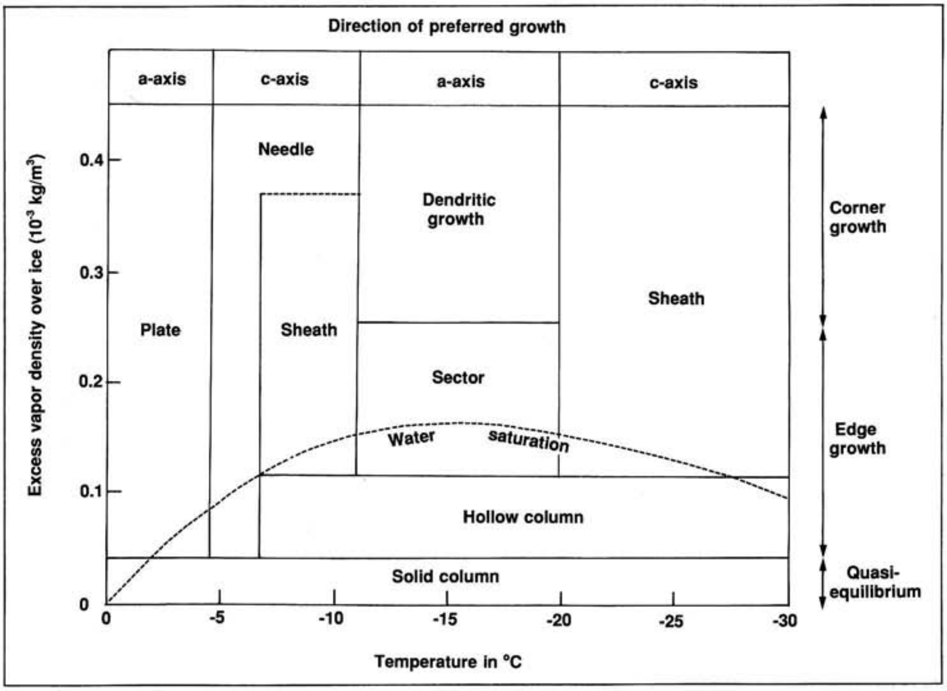


Figure 3.5. Snow crystal growth as a function of excess vapor density (at crystal surface) and temperature. Growth switches between the a- *and* c-*axes depending on temperature, the most important variable. Rounded forms are formed at low vapor density, edges and corners at higher vapor density.*

than in the *c*-axis direction. There is no hexagonal symmetry in the direction of the *c*-axis. Platelike crystals evolve from growth in the direction along the *a*-axes, while needlelike crystals are grown in the direction of the *c*-axis. Regardless of growth direction, snow crystals that grow from the vapor are six-sided due to the influence of basal plane. The rate of growth is also a strong factor in determining which shape a crystal takes. However, it is the excess vapor density (near the surface of crystals) over that which the ice can maintain at the surface that determines the form, along with the temperature. Figure 3.5 shows how growth direction (*a*-axes or *c*-axis) varies to produce the basic crystal forms as a function of temperature for typical atmospheric conditions.

At low excess vapor density, the shapes are basically columns at any temperature. At high growth rates (for example, higher excess vapor densities) growth occurs at edges and corners to produce more complicated crystals, such as dendrites. At these higher growth rates, the complex forms that result are due to transfer of water vapor molecules across the crystal surface. In general, the molecules tend to be deposited at positions where the excess vapor density is highest such as edges and corners. The exact surface processes by which the growth direction switches from *a*-axis to *c*-axis growth as the temperature changes are not well understood. Figure 3.5 clearly shows that temperature is the primary variable determining the crystal form in the atmosphere, and the degree of supersaturation (growth rate) is second in importance.

Some of the complicated forms that arrive on the earth are produced because they pass through different temperature and water vapor density regimes as they pass through the atmosphere. For example, a solid column could be produced in cold air, but on entering a warmer regime, plates might grow on its ends to form a "capped" column.

The rate at which a crystal gains mass determines its size, which in turn depends on the temperature. In general, crystals that have fallen through a cold atmosphere are smaller than those that have been in a warm atmosphere. This is because the thermodynamic processes that govern growth occur faster at warmer temperatures and warmer air can potentially hold more moisture than cold air.

Experienced avalanche observers usually keep a close watch on snow crystals that fall. Observations of snow crystals provide clues about the condition of the atmosphere through which they have fallen. Changes in crystal types during storms, including changes in the amount of riming, can create conditions where one layer does not bond well to the next; this can be of significance in prediction of snow stability. Layers of graupel particles, for example, often do not bond well to their neighbors, which can set up the conditions for snowpack failure. In fact, it has been proposed that the amount of riming can be indirectly related to avalanche formation, including the type of avalanche and the degree of instability. However, at best, this is a second- or third-order effect, which must be integrated with other more important factors in evaluation of snow stability. Also, the breaking of branches of crystals by snow transport at the surface is generally of greater importance than riming for formation of dangerous avalanches. The integrated effects of crystal form, riming, and breakage can all contribute to instability in new snow and its bonding characteristics with old snow layers. However, because such a wide combination of these variables can produce unstable snow, no simple formula is available. Therefore, experienced snow stability analysts concentrate on the integrated effects of these variables as they relate to the mechanics of avalanche formation, rather than an emphasis on any one of these secondary factors in isolation (see Chapter 4).

CLASSIFICATION OF NEWLY FALLEN SNOW CRYSTALS

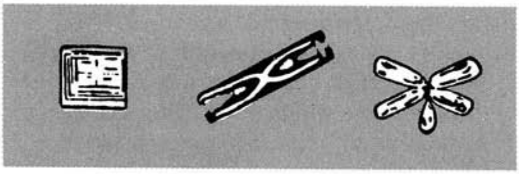
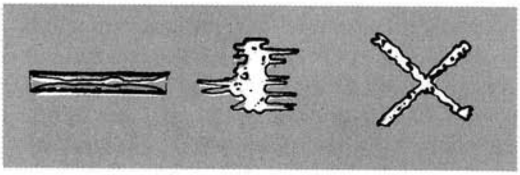
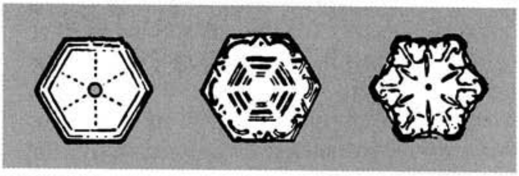
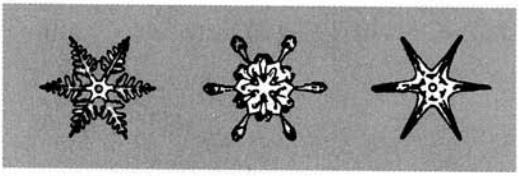
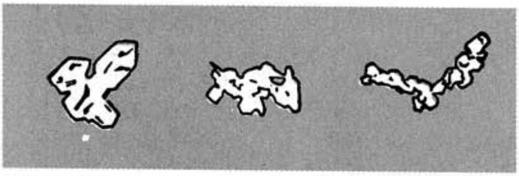
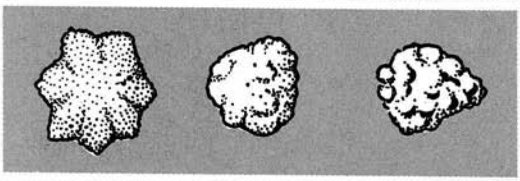
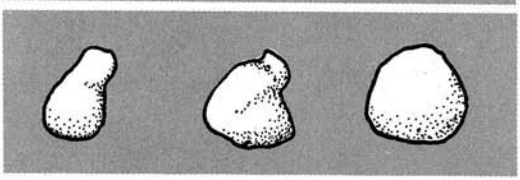
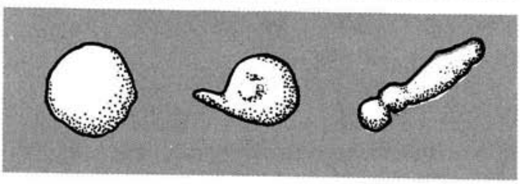
There are three levels of classification for newly fallen crystals depending on the degree of sophistication required in the work. The simplest and most commonly used method is to lump all newly fallen snow into one class with a note about the degree of riming. In most countries, the symbol + is normally used to denote new snow with +r designating rimed newly fallen snow. In Canada and the United States, a third class for a fully rimed group (graupel) with the symbol $\triangle$ is also used.

A more sophisticated system is that of the International Commission on Snow and Ice (ICSI), as shown in Table 3.1. It includes five easily recognizable crystal types as well as a category for irregular crystals and classes for hail and ice pellets (see, for example, Figure 3.6). This system is not as commonly used by avalanche workers as the simple +, r, $\triangle$ system. The reason is that, at best, new crystal forms are a secondary effect in avalanche stability evaluations. In avalanche work, snow is usually analyzed after deposition and a variety of types are mixed together, which complicates a decision. The ICSI system probably represents the most advanced system needed in operational avalanche work. Another even more complex system with 80 categories (including rimed categories) was proposed by Magono and Lee (1966) and is given in Table C.1 in Appendix C. Avalanche workers should be aware of the existence of this complete system, but when snow stability analyses are the objective, one's time may be used more profitably by concentrating on primary (see Chapter 6) rather than secondary effects of precision work in identifying new crystal types.

In classifying snow crystals near the surface of the snowpack, care must be exercised when identifying the predominant crystal form in a sample. It is easy to be distracted by one or two crystals that are easily identifi-

Figure 3.6. Needle and decomposed crystals. (Photo by E. Akitaya)

Table 3.1 ICSI Classification for Newly Fallen Snow Crystals

	**1a** Columns	Short prismatic crystal, solid or hollow	Growth at high supersaturation at -3° to -8°C and below -22°C
	**1b** Needles	Needle-like, approx. cylindrical	Growth at high supersaturation at -3° to -5°C
	**1c** Plates	Plate-like, mostly hexagonal	Growth at high supersaturation at 0° to -3°C and -8° to -25°C
	**1d** Stellar Crystals	Six-fold star-like, planar or spatial	Growth at high supersaturation at temperatures between -12° to -16°C
	**1e** Irregular particles	Clusters of very small crystals	Polycrystals growing at varying environmental conditions
	**1f** Graupel	Heavily rimed particles	Heavy riming of particles by accretion of supercooled water
	**1g** Hail	Laminar internal structure, translucent or milky, glazed surface	Growth by accretion of supercooled water
	**1h** Ice pellets	Transparent, mostly small spheroids	Frozen rain

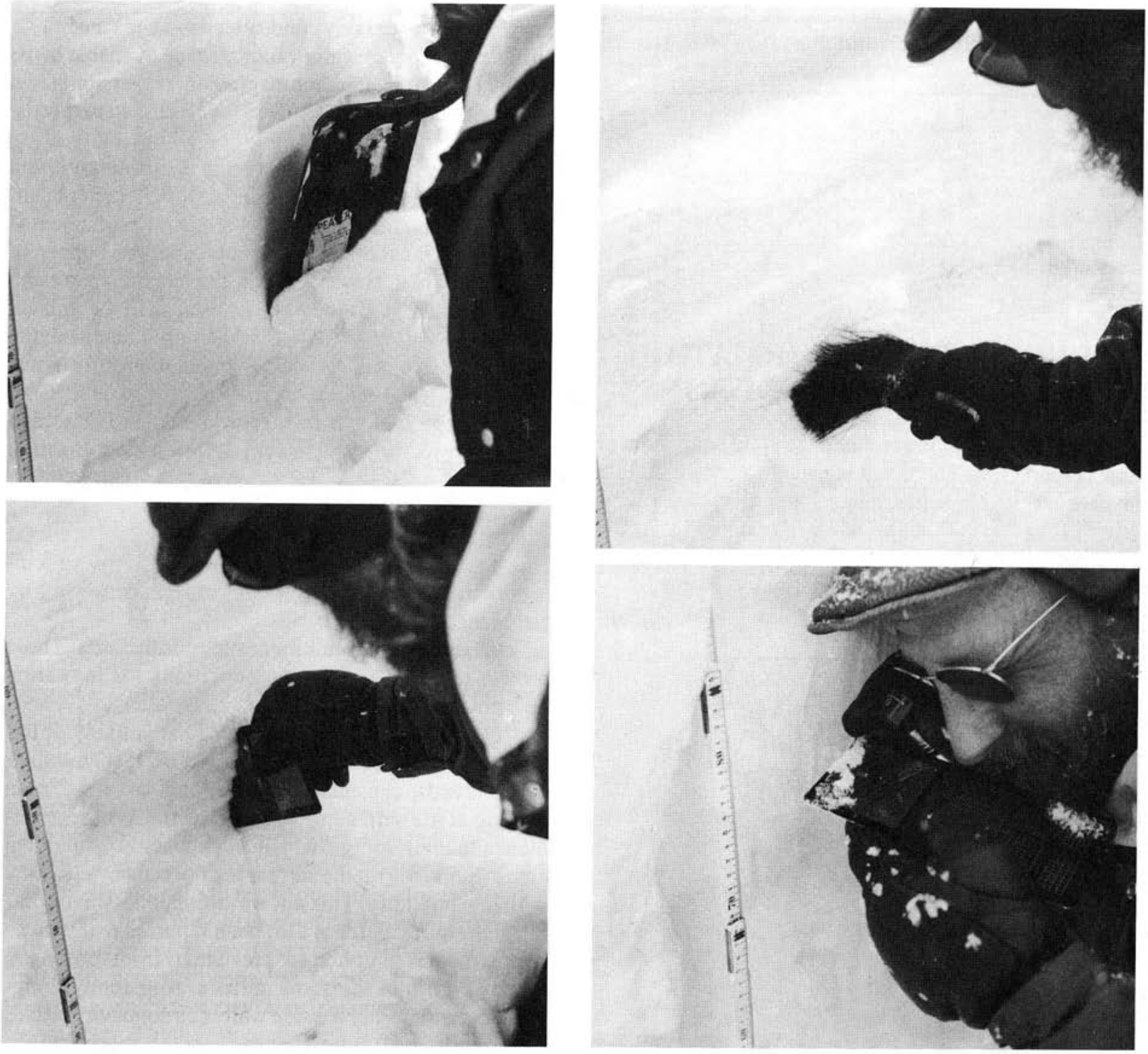


Figure 3.7. Definition of snow layers with shovel, brush, and edge of crystal screen. Use of a hand lens and millimeter crystal grid for examining snow crystals (bottom right). Magnification of 8 to 10X is recommended for field use.

able in a sample rather than the most prevalent form. For normal observations, a pocket or hand lens with about an 8X to 10X power is preferred (Figure 3.7). With higher magnification, the field of view narrows and there is a tendency to focus on the details of individual crystals rather than viewing a good portion of the sample to determine the predominant form. In scientific work, higher-powered lenses and microscopes are sometimes used.

The size of snow crystals is determined in the field by measuring the largest diameter (extension) of the average crystal. Usually avalanche observers quote a range of observed sizes seen on a crystal screen with a millimeter grid, for example, 0.5 to 1 mm. As with crystal forms, crystal size is of secondary importance with respect to avalanche formation. Table 3.2 gives the ICSI recommended size categories.

Table 3.2 ICSI Terms and Sizes for Grain Size Classification

Term	Size (mm)
Very fine	<0.2
Fine	0.2–0.5
Medium	0.5–1.0
Coarse	1.0–2.0
Very coarse	2.0–5.0
Extreme	>5.0

SURFACE HOAR: FORMATION AND GROWTH CONDITIONS

Surface hoar has been termed the solid equivalent of dew (Figure 3.8). The result is usually very weak, thin layers of snow that are extremely important in avalanche release. Surface hoar forms when the water vapor pressure in the air exceeds the equilibrium vapor pressure of ice (snow grains) at the surface. It usually grows rapidly provided two necessary conditions have been met: (1) A sufficient supply of water vapor must be available in the air and (2) a high temperature gradient (inversion) must be present above a snow surface that is chilled below the ice point.

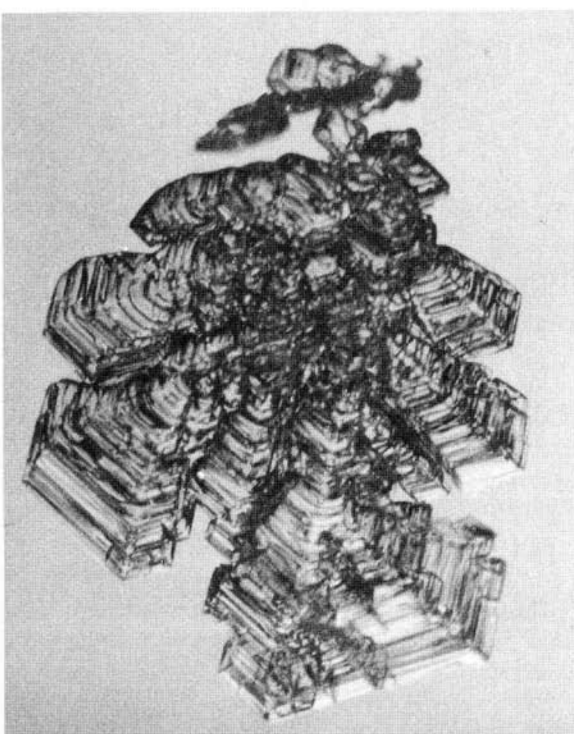


Figure 3.8. Surface hoar. (Photo by E. Akitaya)

Surface hoar usually forms on cold, clear nights with calm or nearly calm conditions in the lowest meter of air. Some investigators believe that slight air movement is necessary near the surface to replenish the supply of vapor deposited. However, if air movement is too rapid (i.e., turbulent), the near-surface air will be mixed, which can destroy the air temperature gradient near the surface. Temperature gradients (inversions) during surface hoar formation are commonly 100° to 300°C/m. Usually the relative humidity in the air is fairly high (>70%) but it is possible for surface hoar to form at lower values if the surface is losing heat by radiating to space. Surface hoar can occur in many forms depending mainly on the temperature. Since it grows from the vapor phase, the dependence of crystal form on temperature applies (Figure 3.5), as has been confirmed by field measurements.

Meteorological conditions favorable for surface hoar growth must set up the two conditions for formation described earlier. If a cold front passes after an overcast day, causing a clear night, surface hoar is likely. It has been reported that even a slight cloud cover, such as high cirrus clouds, can interfere with long-wave radiation cooling at the surface to inhibit growth. Another common situation for growth occurs when supercooled cloud decks (fog) at the surface are overlain by clear sky with low humidity. This allows strong long-wave cooling at the surface. It has also been reported that surface hoar growth is inhibited (or prevented) from forming in concave areas of the snow surface. Apparently, long-wave radiation from the side walls of a concavity strikes the opposite wall instead of escaping to space and the cooling mechanism is less efficient.

Surface hoar can form (and is observed) in any type of climate provided the necessary conditions for growth are met. Since it is extremely fragile, it is also easily destroyed. The agents of destruction include sublimation, wind, surface melt-freeze cycles, and freezing rain. Sometimes the wind destroys surface hoar above the tree line, making more sheltered locations below more hazardous than those near the tops of the mountains.

Buried surface hoar is extremely efficient in producing propagating shear instabilities (fractures) if it is disturbed. Fatal accidents have resulted when skiers on flat terrain precipitated propagating shear fractures in

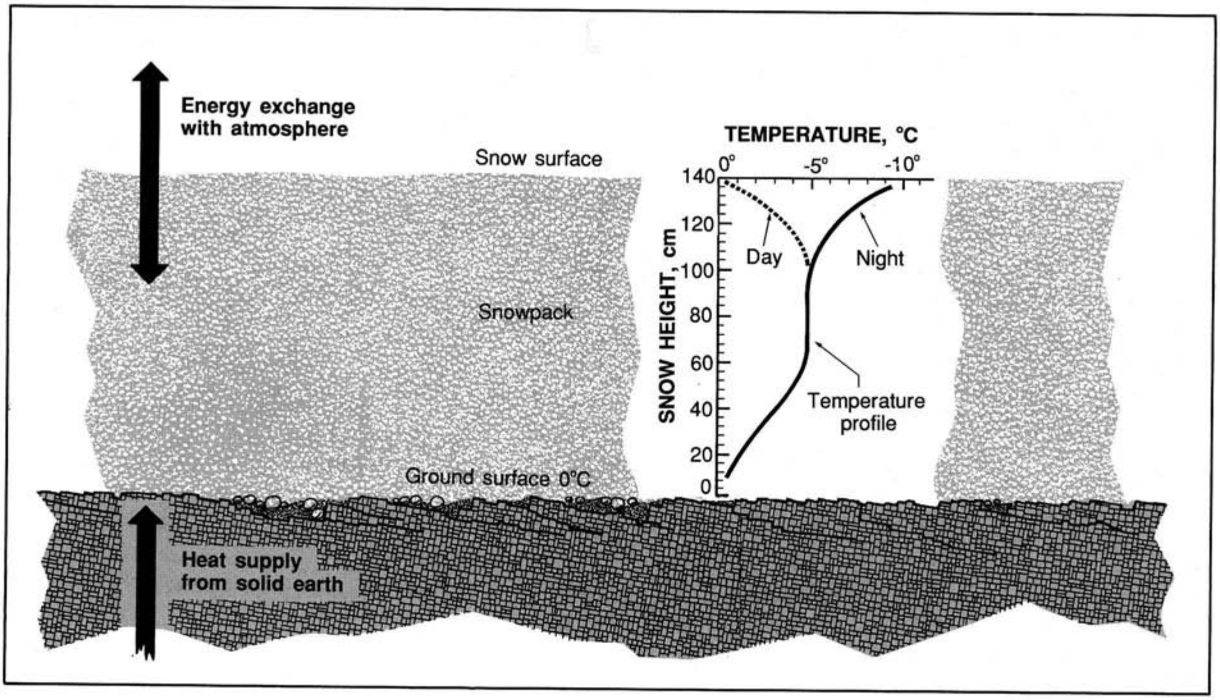


Figure 3.9. Illustration of the temperature variations in a snowpack. Diurnal variations occur in the upper portion. A temperature gradient of 10°C/m is strong enough to produce facets in the snowpack.

surface hoar, which ran upslope to undercut a slab. Such examples show that even though avalanches do not usually form at slope angles below 25° (see Chapter 5), it is not always safe to move on gentle slopes.

Surface hoar may gain strength by bond formation with adjacent layers (see Chapter 4). The persistence of instability depends primarily on the surface hoar layer thickness, which can vary from less than 1 mm to several centimeters. Thick layers can persist for months when buried. Persistence of instability is aided when the crystals are large and unlike their neighbors in form. The typical range of crystal sizes is from less than 1 mm to more than 1 cm in length.

SNOWPACK TEMPERATURES AND TEMPERATURE GRADIENTS

Snowpacks in which avalanches form are bounded by the atmosphere above and the ground surface below. There are circumstances in the high mountains where the lower boundary may be perennial snow or ice as well. Normally, stored heat in the ground from summer warming (most important) and geothermal heat from the earth's center combine to warm the basal layer to 0°C (or close to 0°C). The upper snowpack surface is subjected to cold air during the winter, but snow temperatures at the surface fluctuate wildly in response to daytime and nighttime heating and cooling cycles (called *diurnal fluctuations*) and the prevailing synoptic conditions. Usually these effects combine to produce an upper surface that is cooler (on average) than the lower boundary, which is insulated from the diurnal fluctuations (Figure 3.9).

The long-term effect is a temperature gradient in the snowpack that is a vector quantity having both magnitude and direction. The magnitude of the temperature gradient is defined as the change in temperature (ΔT) divided by the distance (ΔX) over which the temperature changes. By convention, the direction of the temperature gradient is in the direction of increasing temperature (usually downward but sometimes sideways in the snowpack). In metric units, the temperature gradient is expressed in degrees Celsius per meter. When the temperature gradient is 0°C/m throughout, the snowpack is isothermal. The only common occurrence of this is when the snowpack is all at 0°C, which implies a wet snowpack throughout.

In a general sense, temperatures and temperature gradients in snowpacks are coupled together depending on the climate regime. In a maritime climate, prevailing air temperatures are usually mild and snowpacks are deep. These two effects combine to produce weak temperature gradients and warm snow temperatures. In continental climates, snowpacks are shallow and air temperatures are cold to produce strong temperature gradients and cold snow temperatures. The crystal forms that develop under these two scenerios differ, and this has profound effects on the *general* character and timing of avalanches that develop. In a rough, general sense, maritime snowpacks are characterized by relatively strong stable snow and there is a greater bias for avalanches to form in the new snow. Continental snowpacks are usually relatively weak, and often contain buried weak layers in old snow layers, which are susceptible to failures when loaded later. Snow metamorphism (changes in form due to heat flow and pressure), then, largely explains the difference in the character of avalanches in these climates. It must be remembered that crystal forms develop by physical processes (not climate characteristics) so that any crystal form can be found in any mountain range.

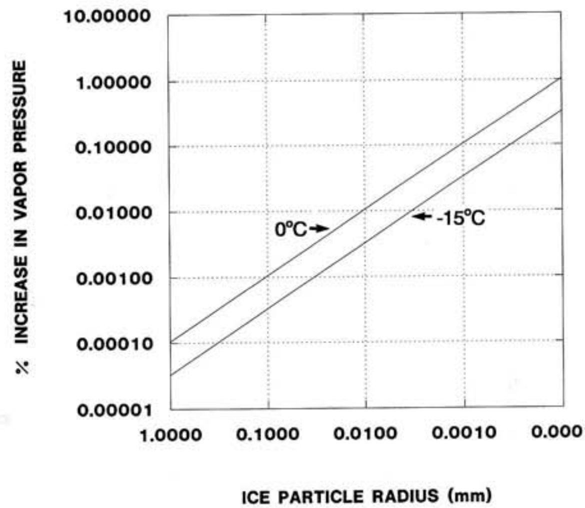


Figure 3.10. Vapor pressure over a curved ice surface as a function of radius of curvature at –15°C and 0°C.

DISAPPEARANCE OF BRANCHES: INITIAL CHANGES IN NEWLY FALLEN SNOW

Once deposited, snow crystals begin to change form immediately. In some cases these initial changes are the direct cause of small avalanches (called *sluffs*). The changes in form also determine future snow strength. Newly fallen snow crystals have grown in an environment that is much more highly supersaturated with water vapor than that encountered once the snow crystals are deposited. Typical values of supersaturation in the atmosphere can be several tens of percent while they are typically less than 1% inside a snowpack. New snow crystals are unstable; they would need the highly supersaturated environment they came from to continue growth in the same form or to maintain their forms. In general, a large ratio of crystal surface area to volume is unstable once the atmospheric growth stops. In the snowpack, snow crystals with the largest surface-to-volume ratio (such as dendrites) are the most unstable and change form most quickly. The minimum surface-to-volume ratio form is theoretically a sphere, therefore rounded graupel particles are very stable (they persist for long periods of time).

The physical reason that the intricate branches of snow crystals initially disappear is because the vapor pressure over sharply curved branch features is very high (the pressure varies inversely with the radius of curvature of the surface). The vapor pressure is higher over a convex surface than it is over a concave surface. Thus, sharp branches promote sublimation by loss of molecules from the ice surface into the surrounding air.

The initial disappearance of branches by curvature does not last long. A key concept with respect to changes in form in dry snow is illustrated by comparing the vapor pressure changes due to curvature with the vapor pressure differences in the pore space as the temperature varies (for example, due to an applied temperature gradient; see Figure 3.9). The pressure changes due to curvature effects are very small. For example, the pressure due to curvature effects increases by about 0.03% for a branch point of radius 10^{-3} mm over that for a flat surface at 0°C and the increase is about 0.1% at –15°C. By contrast, the saturation vapor pressure (with respect to ice) in the pore space of a snowpack increases by more than 300% as air temperature increases from –15° to 0°C (Figure 3.10). This shows why overall temperature differences in dry snow due to snowpack temperature

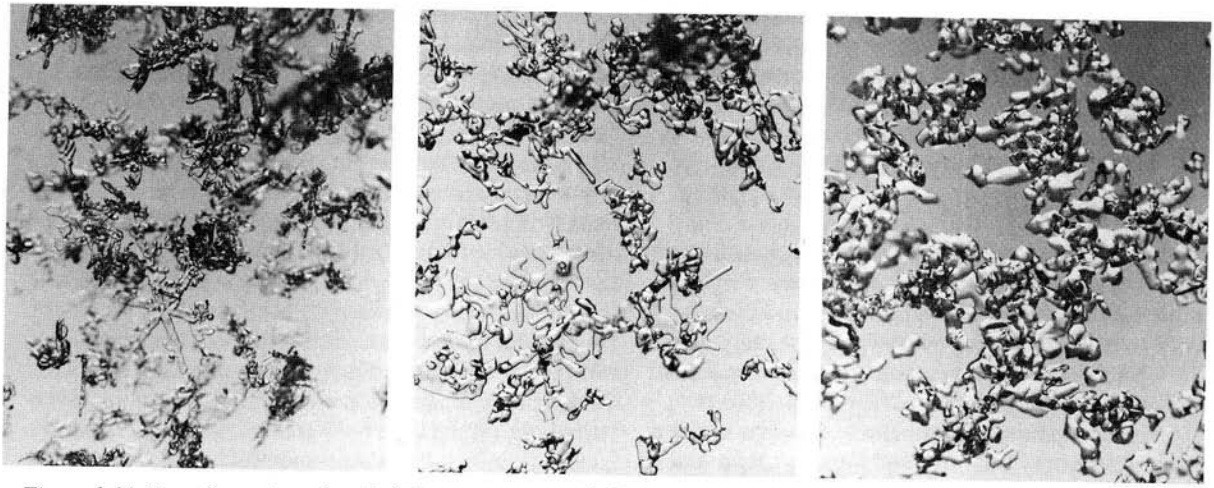


Figure 3.11. Transformation of newly fallen snow to rounded forms in three stages. (Photos by E. LaChapelle)

gradients provide the driving mechanism for snow metamorphism rather than curvature effects. Both grain curvature and overburden pressure due to the weight of overlying snow layers will also influence dry snow metamorphism because they are always acting (Figure 3.11).

The surface regions of the snowpack are those in which the temperature gradient is usually highest. In the general case, the metamorphism will be influenced by both the vapor pressure gradient (caused by the snowpack temperature gradient) in the pore space as well as curvature effects. However, the gradient effect usually drives the metamorphism in a snowpack unless the radius of curvature on branch surfaces is less than about 10^{-2} mm. Experimental data show that it takes more than 10 times as long for a dendritic crystal to decompose to a rounded form in a laboratory under constant temperature conditions than in the field where there is always a temperature gradient (see Figure 3.12). Curvature effects are important in that sharp branch points are preferential sites (high pressure) for sublimation to occur even with an applied temperature gradient. Curvature effects are therefore necessary to explain rounding of complex forms with a high surface-to-volume ratio. However, in general, the temperature and temperature gradient determine the rate of metamorphism of dry snow for either newly fallen snow or snow found at depth in the snowpack once initial rounding has taken place.

When branches disappear, the result is an initial decrease in the average particle size. This process has been called *destructive metamorphism*. However, in the snowpack, size begins to increase after the branches disappear. Sublimation supplies water vapor to increase the vapor pressure in the vicinity of the small particles or branches. Once the water vapor moves (due to the applied snowpack temperature gradient), it tends to condense on larger particles where the water vapor pressure is lower or vapor density is lower. This sublimation (condensation mechanism) encourages growth

Figure 3.12. Sketch of crystal metamorphism (at constant temperature) by curvature effects. The numbers give the time in days. (After Bader, 1939)

of larger particles at the expense of the small ones. By this means (and perhaps heat flow paths within the structure), the average particle size eventually slowly increases when a mixture of sizes is present.

DRY SNOW METAMORPHISM IN THE SEASONAL SNOW COVER

Inside the snowpack the variety of forms is very limited because they are insulated by their neighbors and because the physical conditions around them vary slowly. In this section, the physical processes and changing crystal forms found within the alpine snowpack are described.

The term *metamorphism* includes changes in form due to temperature (heat flow) and overburden pressure. However, in seasonal snow in which avalanches form, the changes in crystal forms are due almost entirely to heat flow within the snowpack. Overburden pressure densifies alpine snow by rearranging the grains (see Chapter 4). It has a role in accelerating metamorphism (forming rounded grains). Overburden pressure *dominates* changes in form only in snow that has densified beyond the normal limits found for seasonal snow, such as firn snow on glaciers that is more than one year old.

A major difference between snow crystal formation in the atmosphere and snow metamorphism in the seasonal alpine snowpack is the degree of supersaturation in the air surrounding the crystals. In the alpine snowpack, the air in the pores is slightly supersaturated with values less than 1% being typical. In the atmosphere, supersaturation with respect to ice can vary from near 0% to 50% or more. For snow crystals forming in the atmosphere, temperature is the primary factor that determines the crystal form—the amount of supersaturation is secondary. For the alpine snowpack, the temperature gradient is the most important factor. There, heat flow and crystal growth rate are most strongly related to vapor diffusion through the pore space. This diffusion process is controlled by the temperature gradient.

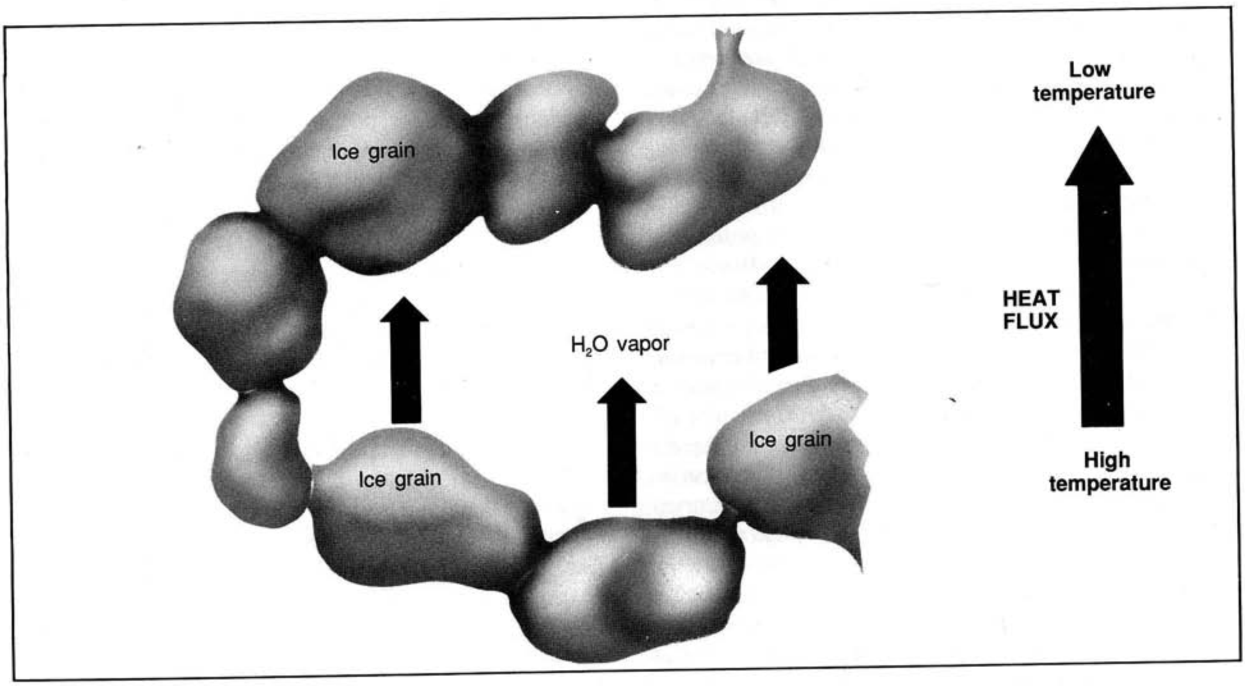


Figure 3.13. Heat flux (carried by the water vapor) occurs from the bottom to the top of the snowpack when temperatures increase with depth from the surface. Water vapor deposits on the underside of crystals and leaves the top in a "hand-to-hand" process.

The vapor diffusion process operates as follows. In a typical dry alpine snowpack, the temperature is near 0°C at the bottom due to stored heat from summer and geothermal heating. The snow temperature decreases toward the snow surface, which is exposed to the cold winter air. Since saturated warm air can hold more vapor than saturated cold air, a higher overall water vapor pressure exists in the pores of the snow at the bottom of the snowpack, and the water vapor is forced to move up. Water vapor moves up through the pore space by leaving the top of one crystal and condensing on a crystal somewhere above. Thus, the motion and heat flow are upward by a hand-to-hand process in dry snow. The rate of motion determines the crystal forms that develop under this recrystallization process (Figure 3.13).

CRYSTAL FORMS IN DRY SEASONAL ALPINE SNOW

For the low supersaturations found in the seasonal snow cover, a strong correlation exists between the overall growth rate and the crystal forms. (The terms *crystal* and *grain* are used interchangeably in this book to denote single ice particles.) The growth rate and crystal forms in a seasonal snow cover depend on three important variables: (1) temperature gradient, (2) the temperature, and (3) the pore space size. Of these three variables, the snowpack temperature gradient is the most important in determining the crystal form in the alpine snowpack. Crystal metamorphism in snowpacks is driven by the vapor density gradient therein. Since the equilibrium vapor pressure over an ice surface is a *nonlinear* function of temperature and temperature gradient, the discussion in this section immediately follows. The discussion here is based on temperature gradients (rather than vapor pressure gradients) since temperatures are normally measured by avalanche workers. In general, the highest crystal growth rates are found for large temperature gradients, high temperatures, and large spaces between the crystals. These conditions produce angular or faceted grains, which may later develop steps and striations on their surfaces, ultimately resulting in large cup crystals called *depth hoar.*

At the other extreme, low growth rates produce rounded forms. The lowest rates would occur with low temperature gradients, tightly packed crystals, and *lower* temperatures. In alpine snow covers, low growth rates are almost always associated with low temperature gradients and *high* temperatures because the temperature and temperature gradient are usually coupled in an average sense: Low temperature gradients and high temperatures usually occur together. The fact that both highly faceted crystals (depth hoar) and well-rounded crystals are found at high snowpack temperatures shows that the temperature gradient is more important than the temperature in controlling the crystal forms in alpine snow. The critical temperature gradient to produce faceted forms in alpine snow is about 10°C/m; below this value rounded forms tend to appear (Figures 3.14, 3.15, and 3.16).

As the temperature decreases in alpine snow, the degree of supersaturation adjacent to crystal surfaces increases rapidly. This should promote formation of facets and corners on snow grains similar to crystal formation in the atmosphere (see the right vertical scale of Figure 3.5). However, at lower temperatures, the growth rate must slow down. Therefore, in cold snow, facets form slowly; apparently under most conditions the growth rate is high enough to form large, cup-shaped (depth hoar) crystals only near the ground (where the temperatures are usually high).

Therefore, the unifying concept in alpine snow is the growth rate: At high growth rates, faceted crystals form; at slow growth rates, rounded crystals form. Higher growth rates imply high temperature gradients, larger crystals, and large pore spaces with the highest growth rates occurring at high temperatures (at the bottom of the snowpack). Lower growth rates imply lower temperature gradients (which imply high temperatures in an alpine snowpack) and small pore spaces.

In general, crystals that are produced under high growth rates form weak, unstable snow that is often responsible for serious avalanche conditions. Examples include surface hoar, radiation recrystallization, faceted snow, and depth hoar. The majority of these types form in cold climates characterized by high temperature gradients and a persistence of instability due to the cold. The class of crystals formed at high growth rates accounts for big avalanches and large numbers of fatalities. Backcountry travelers must pay close attention when such crystals are present in the snow cover. Unfortunately, the growth rate under which crystals are produced is not measured in avalanche work. Instead, avalanche workers look for the classes of crystals produced at rapid rates or they measure temperature gradients as a guide. High growth rate crystals form preferentially in cold, continental-type climates but not exclusively. In the United States, there is a mix of snow climates and people exposure but it comes as no surprise that the greatest

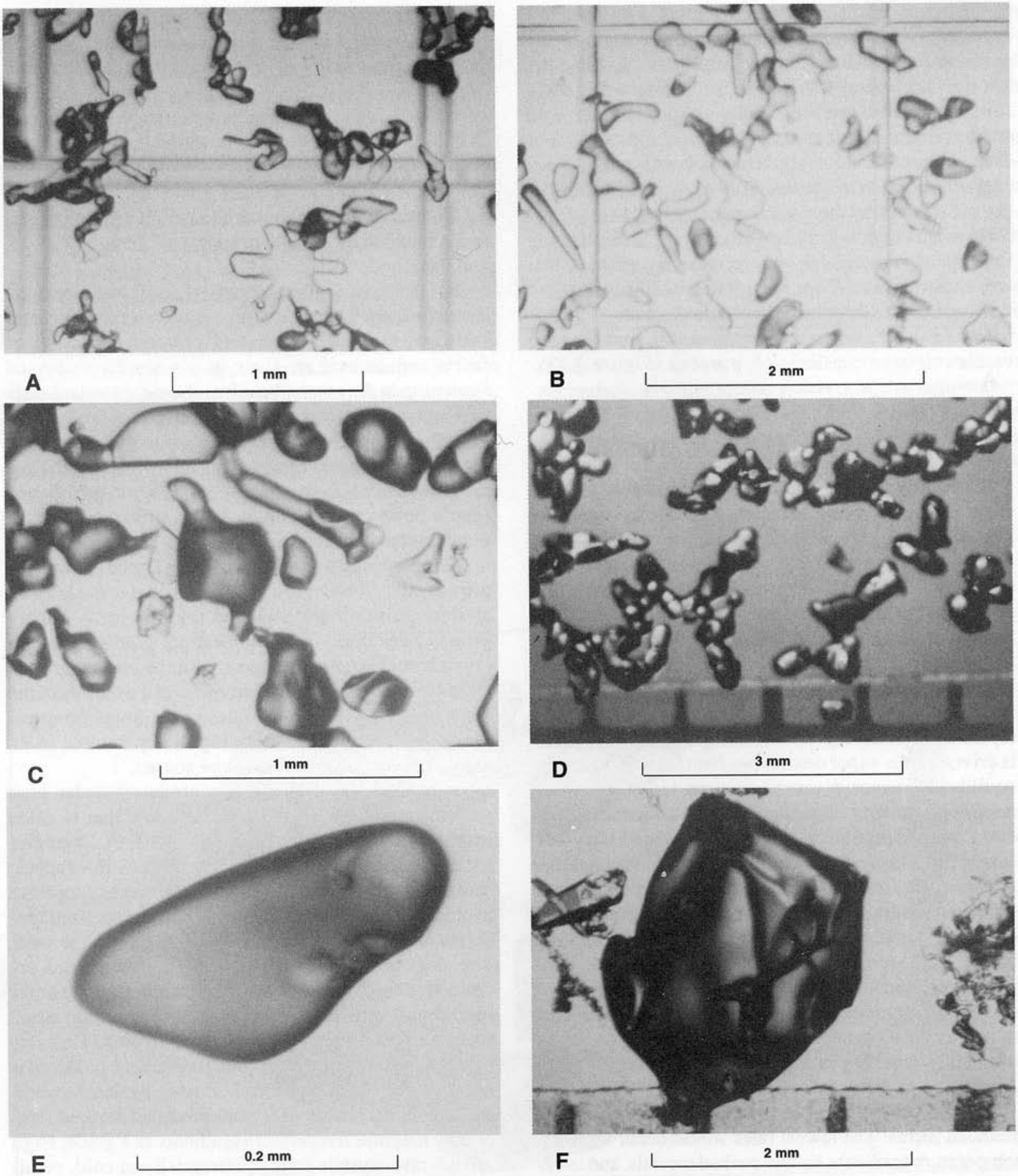


Figure 3.14. A: Initial metamorphism, original forms still visible; B: decomposing forms; C: development of rounded forms; D: rounded forms with bond formation; E: close-up of rounded form; F: faceted form with new snow particles. (Photos by R. Perla)

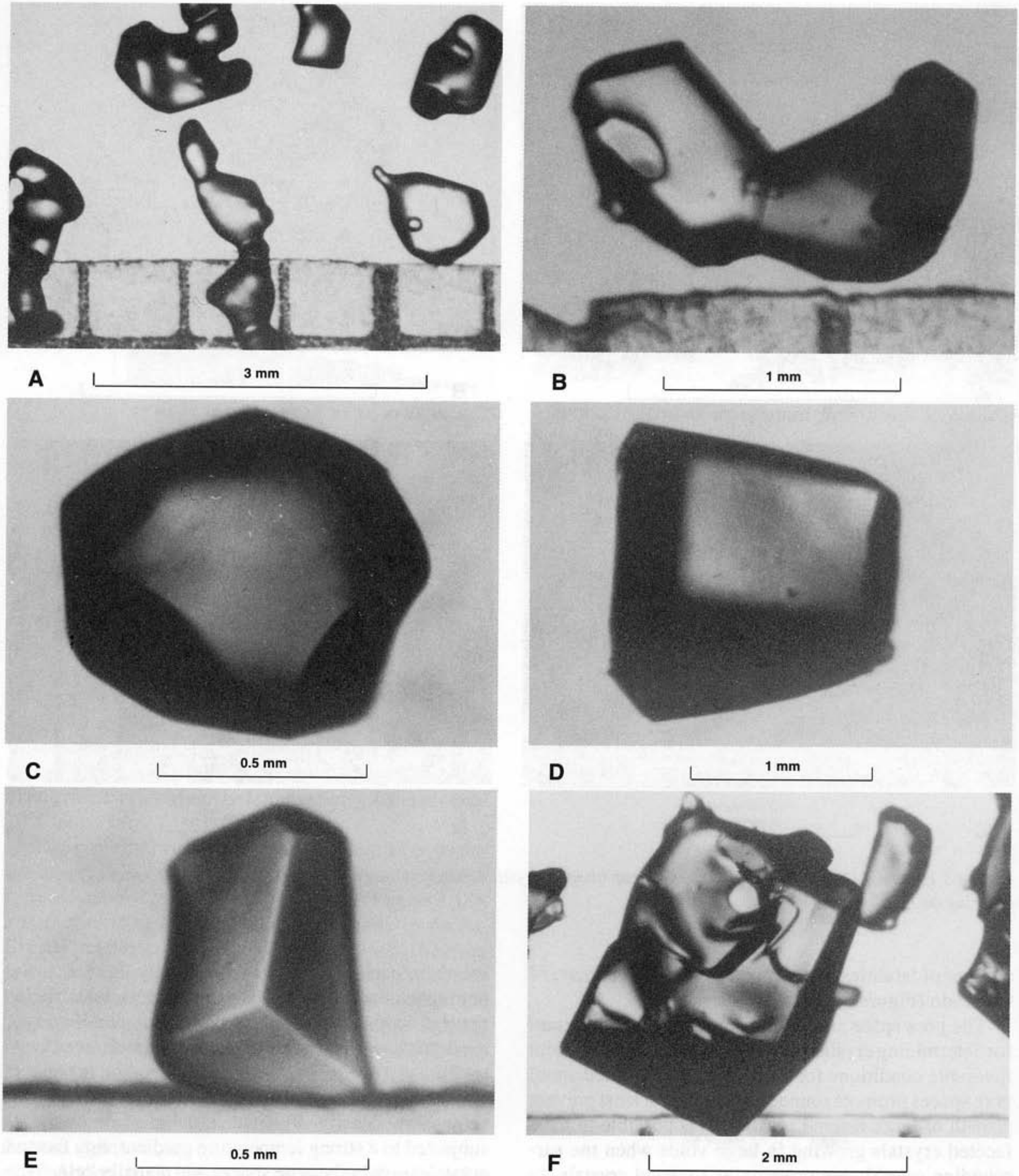


Figure 3.15. A: Faceted forms that have undergone rounding; B–F: examples of faceted forms. (Photos by R. Perla)

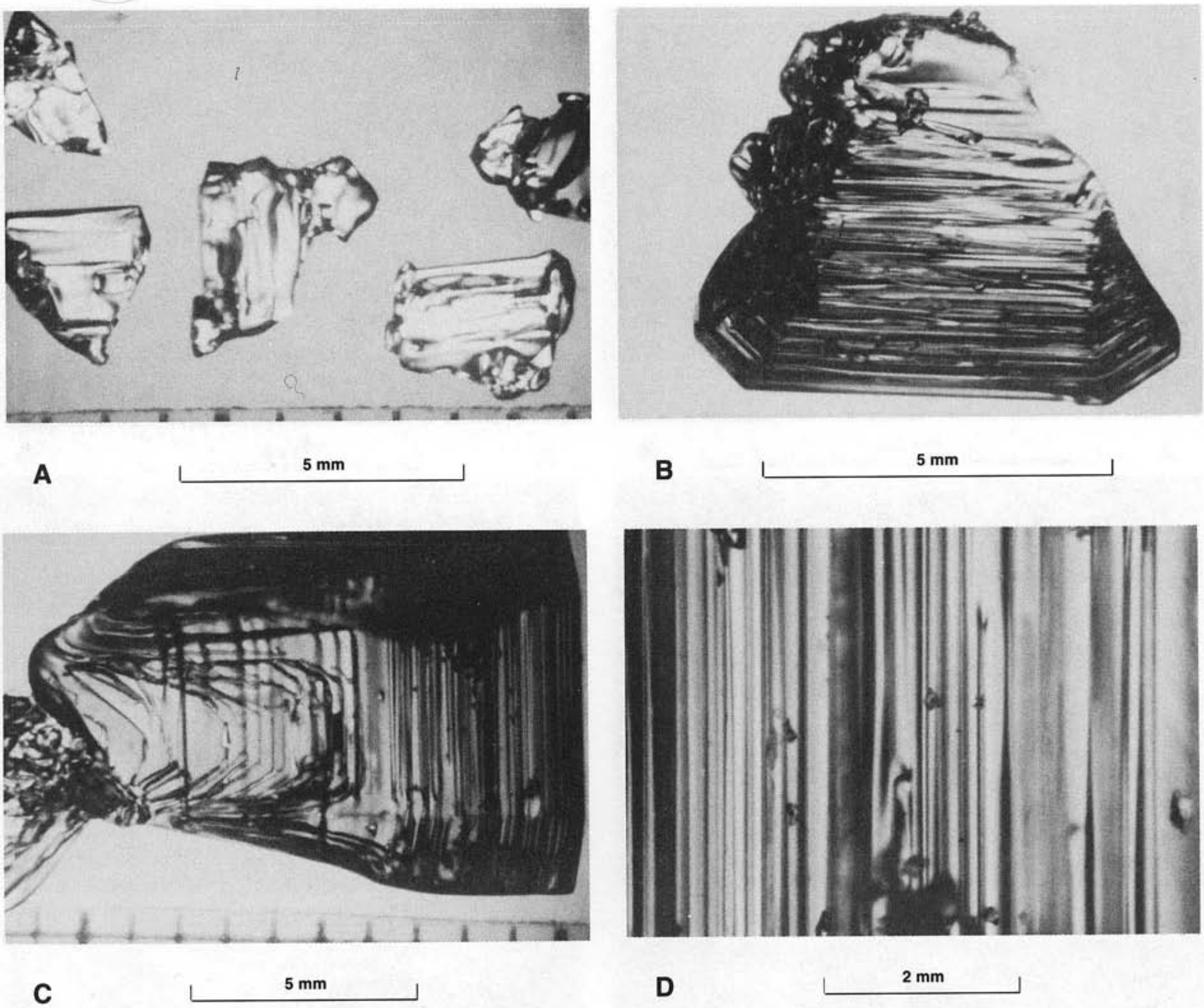


Figure 3.16. A–C: Depth hoar and hollow or cup-shaped crystals. Note the large sizes of crystals. D: Close-up of steps and layering on depth hoar. (Photos by R. Perla)

number of fatalities occurs in the continental climate of Colorado (Figure 3.17).

The pore space size and geometry is also important for determining crystal forms. Large pore spaces present favorable conditions for formation of facets and small pore spaces promote rounded crystals or at least prevent growth of large faceted crystals. It is possible to have faceted crystals growing in large voids when the surrounding crystals are growing as rounded crystals. In climates where depth hoar grows at the bottom of snowpacks, it is common to bootpack the slopes in early season to pack the crystals more tightly. In reality, the pore space size is decreased by this process, which helps prevent depth hoar formation thereafter and, of course, the depth hoar present is compressed as well (see Chapter 9).

Depth hoar is not always weak. When fine-grained snow with density of about 350 kg/m^3 or higher is subjected to a strong temperature gradient, tiny faceted crystals form in the pore spaces and actually "glue" the grains together to increase strength and hardness. It has also been observed that recrystallization in the presence

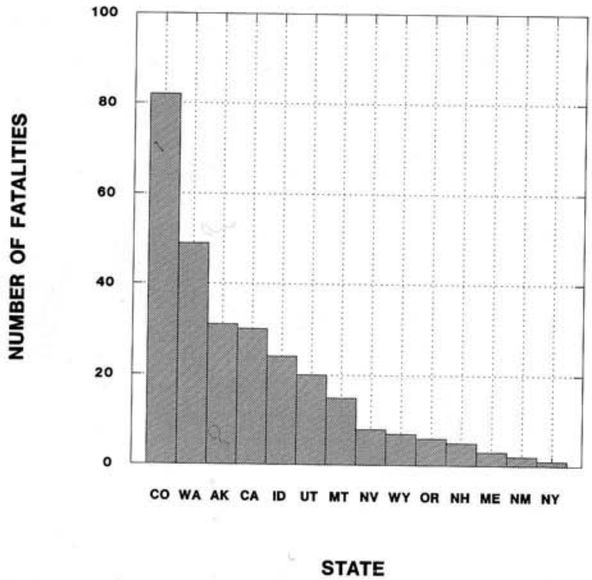


Figure 3.17. Thirty-five years of U.S. avalanche fatalities by state. Colorado has a continental snow climate that allows for the persistence of buried structural weaknesses.

of strong temperature gradients appears to proceed more rapidly at high altitudes: Lower average air density in the pore space may allow more rapid diffusion of water vapor.

Mixed forms include rounded crystals with facets or faceted crystals with rounding; they develop at intermediate growth rates when the temperature gradient is near 10°C/m.

The preferred name for avalanche workers to describe crystals developing under the slowest growth rates is *rounded forms* or simply *rounds* (Figure 3.18). For faster rates, the term is *faceted forms* or *facets.* Crystal scientists also call these forms *equilibrium forms* and *kinetic forms,* respectively. However, laboratory studies of single-crystal growth show that the "equilibrium" form is actually a hexagonal (edged) plate at cold temperatures (below about –10°C) and rounded at higher temperatures. Therefore, it is felt that the term *rounded crystals* is a better descriptive label for the slow growth rate forms observed at warm temperatures in alpine snow.

In Appendix C, Table C.2 gives classifications for *dry* seasonal snow from the ICSI snow classification system including surface forms and crusts. Two levels

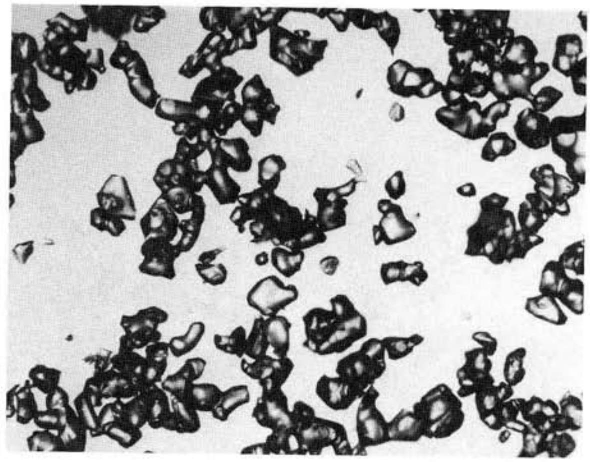


Figure 3.18. Grains transformed from facets to rounded forms. (Photo by E. Akitaya)

of classification are given. The basic system (first column) has classes for dry snow with three additional ones for surface forms and crusts—feathery (hoar) crystals, ice masses, and crusts—and one class for wet snow (Figure 3.19). The basic system is recommended for avalanche workers in snow profile work and is shown here in Table 3.3. The subclasses in column 3 of Table

Table 3.3 Basic ICSI system for classifying snow crystals

Decomposing and Fragmented Precipitation Particles	/
Rounded Grains (Monocrystals)	●
Faceted Crystals	□
Cup-Shaped Crystals; Depth Hoar	∧
Wet Grains	○
Feathery Crystals	∨
Ice Masses	▬
Surface Deposits and Crusts	

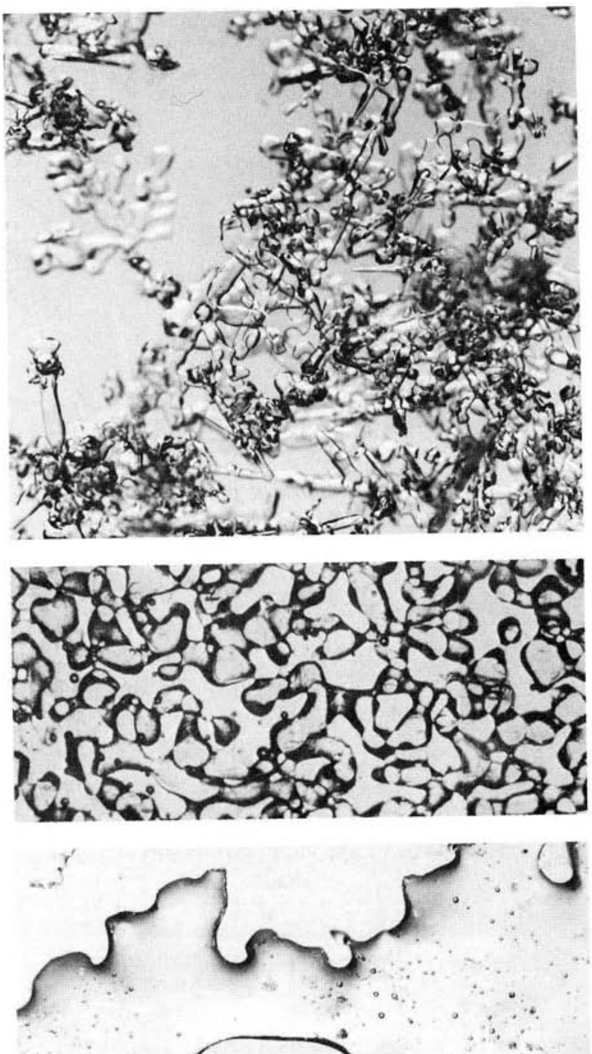


Figure 3.19. Thin sections of snow illustrating newly fallen snow, old dense snow (rounded forms), and a thin ice crust. (Photos by E. LaChapelle)

C.2 are also used by avalanche workers in some cases; sometimes a mixture of basic and subclasses is used. The level of complexity in the subclasses may not be necessary in avalanche work since crystal forms are usually a second-order effect in avalanche formation. Avalanche workers normally use the pictorial (graphical) symbols, but numerical and letter symbols are usually avoided except in computer plot packages.

In popular literature originating mostly from the United States, the term *equitemperature metamorphism* has been used to describe the process of generating rounded forms, and *temperature gradient metamorphism* has been used to describe development of faceted forms. Since temperature gradients largely determine the rate of growth, and therefore the form, in either case and since "equitemperature" conditions are nearly impossible to have in dry alpine snow, these terms are not recommended.

In general, the size of crystals found in the interior of a dry snowpack depends on their age and the physical conditions determining the growth rate at their location. The largest crystals are often depth hoar because they have been present the longest and they grow at the highest rates. The smallest crystals are usually found where growth is the slowest due to a low imposed temperature gradient and tightly packed condition (small pore spaces), for example, from wind effects creating small, fine-grained snow. The size of snow crystals is an important element in snow crystal identification in the field. Experienced observers often try to determine the size of crystals first when trying to identify crystals, because the size gives an important clue about the growth rate and hence the expected crystal form.

METAMORPHISM OF WET SNOW

When snow becomes wet, conditions with respect to heat flow and subsequent metamorphism change greatly. An added complexity is that wet snow can consist of air, ice crystals, and water (a three-phase system). Contrary to what might be thought at first glance, wet snow is not exactly at a temperature of 0°C. In fact, small differences in temperature within wet snow are responsible for the metamorphism of wet snow. Just as in dry snow, the pressure over the curved ice surfaces in wet snow particles is inversely proportional to the radius of curvature of the surface. This causes the melting temperatures of the particles to be size dependent—small particles have a lower melting temperature than large ones. The mechanism of metamorphism changes in wet snow as the mix of water, air, and ice contents changes.

SNOW WITH HIGH WATER CONTENT

Water-saturated snow is responsible for slush avalanches in the Arctic and it is commonly found at impermeable layers (including the ground) when large avalanches release due to water lubrication.

In water-saturated snow (slush), the particles are usually entirely separated from each other by water (Table 3.4). This is known to occur when the water content exceeds approximately 15% by bulk volume. Since small particles have a lower melting temperature than large ones, they melt first. The heat of melting comes from the larger particles, which undergo surface refreezing (release of heat) and an increase in size.

In water-saturated snow, the metamorphism or particle growth is caused by the heat flux through the water during the melting–refreezing process. The rate of growth is very rapid for any distribution of particles that includes sizes below about 1 mm. The rate of growth (metamorphism) decreases and the average size of the particles increases with time. Since metamorphism proceeds as a result of temperature differences between particles, due to differences in size, theoretically metamorphism should stop if all particles were *spheres of the same size,* and the particles would slowly melt. However, such a situation would be virtually impossible to achieve in alpine snow and experimental evidence suggests that particles usually attain elliptical shapes.

The differences in melting temperature due to curvature of ice particles are very small in wet snow. It is estimated that the melting temperature of a 1-mm-radius particle is 10^{-4} °C below freezing and that for a particle 0.1 mm in size is 10^{-3} °C below freezing. However, due to the ease with which heat is transferred (by liquid water) among neighbors when the melting–refreezing process takes place, the cannibalism in which small particles melt to produce larger particles is very rapid in spite of these small temperature differences (Figure 3.20).

The term *very wet snow* (or funicular regime) is applied to snow with an intermediate water content (8% to 15%). The presence of air bubbles in this case is important in retarding the growth of particles in wet snow. Since heat conduction is much slower in air than water, the main effect of air bubbles is to reduce the area available for heat conduction through the water and slow grain growth.

Table 3.4 ICSI classification system for water content of snow

Term	Remarks	Water Content (% by volume)	Graphic Symbol
Dry	Usually T is below 0°C, but dry snow can occur at any temperature up to 0°C. Disaggregated snow grains have little tendency to adhere to each other when pressed together, as in making a snowball.	0%	
Moist	T = 0°C. The water is not visible even at 10 x magnification. When lightly crushed, the snow has a distinct tendency to stick together.	< 3%	
Wet	T = 0°C. The water can be recognized at 10 x magnification by its meniscus between adjacent snow grains, but water cannot be pressed out by moderately squeezing the snow in the hands. (Pendular regime)	3-8%	
Very Wet	T = 0°C. The water can be pressed out by moderately squeezing the snow in the hands, but there is an appreciable amount of air confined within the pores. (Funicular regime)	8-15%	
Slush	T = 0°C. The snow is flooded with water and contains a relatively small amount of air.	> 15%	

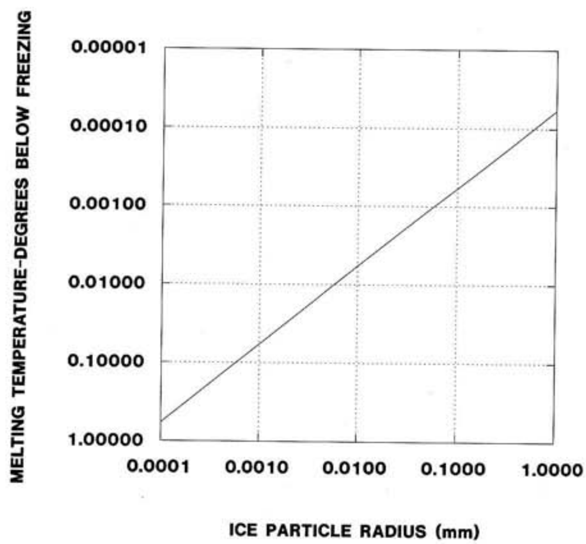


Figure 3.20. Melting temperature of ice particles in water as a function of particle radius.

SNOW WITH LOW WATER CONTENT

When the water content drops below about 8%, grain growth occurs by vapor flux through the air in the pores. For this case, the air in the pore space is completely interconnected. The vapor flux responsible for metamorphism is caused by the vapor pressure excess over the convex portion of the grain due to curvature: The vapor pressure increases as the radius of curvature decreases just as in dry snow. The resulting grain growth is very slow. For water contents between 3% and 8% by volume, snow is termed *wet* although water cannot be pressed by squeezing gently by hand. However, a meniscus of water may be seen between grains with a hand lens. The border between wet and moist snow (less than 3% water content) can be difficult to recognize in the field.

As the water content in wet snow decreases, a capillary pressure develops and the pressure increases in the pore water between the grains (Figure 3.21). This pressure is the difference between the air and water pressures, and it forces water out of the pore space between the grains. When the water content becomes low enough (usually less than about 7% by volume), the grains form ice-to-ice bonds and they readily form clusters. (Since alpine snow is highly porous and therefore permeable to water flow, water usually drains freely through it unless an impermeable layer is present nearby.) Low water content snow is a common occurrence and the ice bonds between the grains can give the snow some strength. The distinguishing feature is that the grains cluster together and this is easily recognized in the field. Snow that clusters is called *moist*—water is not visible even with a hand lens (water content of less than 3%), but the snow should stick together when gently pressed to form a snowball (Figures 3.22 and 3.23). If there is any doubt about whether snow is wet at low water contents the snow temperature should be measured. If the temperature is 0°C, the snow has some degree of wetness. The easiest and best method of directly measuring snow water content is by measuring the dielectric constant of

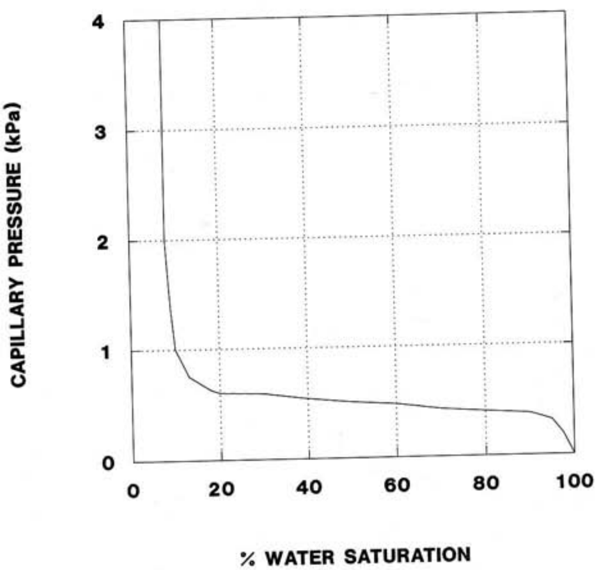


Figure 3.21. Capillary pressure in wet snow as a function of water content (% saturation).

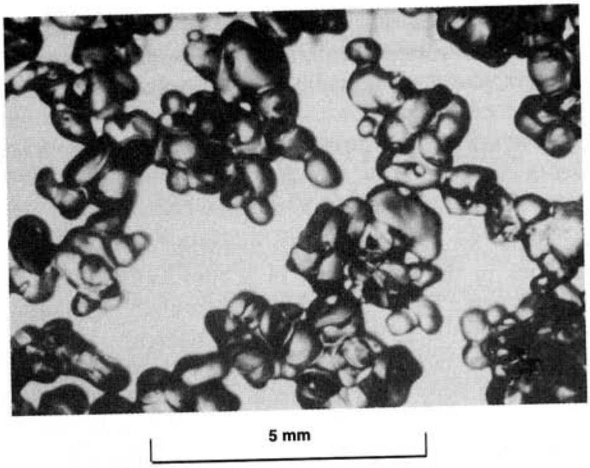


Figure 3.22. Clusters of wet snow (frozen). (Photo by R. Perla)

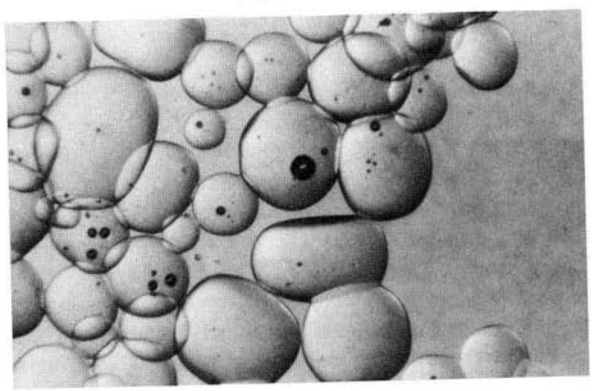


Figure 3.23. Slush. (Photo by S. Colbeck)

snow (a function of density and water content) using a portable instrument. The disadvantage is the expense of the instrument.

In general, growth in wet snow speeds up with increasing water content, while the strength of wet snow decreases with increasing water content. The transition point where grain growth occurs mostly by vapor flux occurs when 8% or less of the pore volume is filled by water. The term *pendular regime* has been applied for this low water content snow, but descriptive terms based on an approximate measure of the actual water content are used by avalanche field workers. Table 3.4 lists the descriptive terms from the ICSI classification system.

CLASSIFICATION OF WET SNOW

In Appendix C, Table C.3 gives the ICSI classification for wet snow. The basic scheme has only one class and three subclasses are listed. In avalanche work, the basic class and the subclasses are sometimes used. The classes in Table C.3 refer only to snow that has been wetted long enough to produce rounded forms. For example, if faceted snow suddenly becomes wet, it is still classed as faceted according to its morphology.

BOND FORMATION IN DRY ALPINE SNOW

Formation of bonds (also called *sintering*) between snow crystals is a crucial element in avalanche formation since bond formation is closely related to snow strength. Formation of bonds occurs by diffusion of water vapor through the pore space between grains as well as by molecular motion on the surfaces when crystals touch. Therefore, like metamorphism, bond formation is strongly affected by conditions at and near crystal and grain surfaces (Figure 3.24).

Although curvature effects are *not* generally the primary driving force to move the water vapor through the pore space in an alpine snowpack, they can play a role in determining where water vapor molecules condense on a crystal to form a bond. In general, the vapor pressure is higher over convex surfaces than flat surfaces or concave surfaces. Thus, there is a greater tendency for water vapor to condense at concave surfaces (low-pressure areas) where grains touch to form necks. Laboratory experiments performed at constant temperature (a virtual impossibility in a dry snowpack) showed that about 90% of the mass deposition at necks

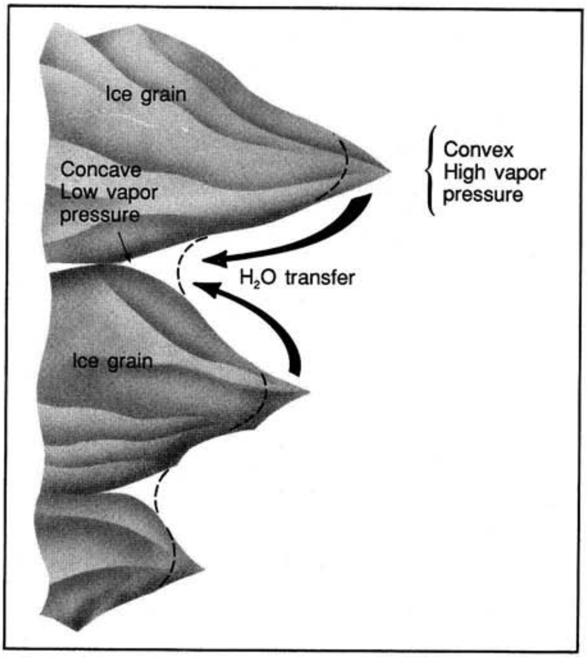


Figure 3.24. Vapor pressure is higher over a convex ice surface than over a concave ice surface. This aids mass transfer to concave surfaces.

between spherical particles occurs by vapor transport through the air from convex regions to the (concave) neck and about 10% occurs by motion of molecules over the surface toward the neck (Figures 3.25 and 3.26).

As in metamorphism, the temperature gradient and pore space geometry in a snowpack strongly affect the character of bond formation. Since thermodynamic processes occur faster at warmer temperatures, bond formation increases rapidly with increasing snow temperature. This concept is very important in avalanche stability evaluation. The temperature also affects the conditions of the surface of the crystals. There is a highly mobile "liquidlike layer" at the surface of crystals that thickens with increasing temperature. One idea proposed is that this layer can partially obscure the potential landing sites on the surface (ice lattice features) at high temperatures so that the landing sites for molecules are controlled more by curvature effects as long as the overall growth rate is not too high. This combination produces rounded crystals with strong bonds at the necks.

When the temperature gradient is high, mass transport is rapid and the vapor pressure gradient that provides the driving force is apparently high enough such

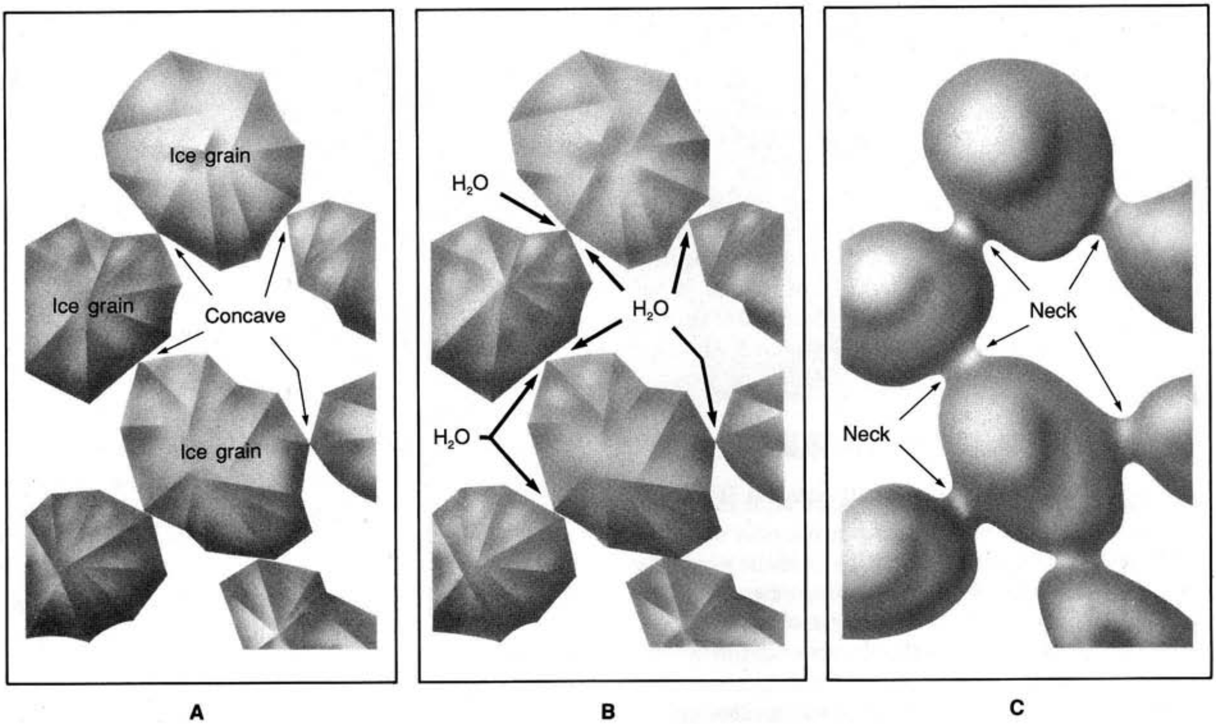


Figure 3.25. Bond formation in dry snow. Water molecules move through the air and along the grain surfaces to concave regions (low vapor pressure) to form bonds (necks) in a process called sintering.

that curvature effects do not have much influence on the landing sites for water vapor molecules. In this case, observations show that water vapor molecules are no longer deposited preferentially at necks, and growth occurs on the faces and edges of crystals as well. The result is weak snow with faceted grains and anisotropic character. For example, depth hoar develops in this manner and it is relatively weaker in shear than in compression (see Chapter 4). Depth hoar can be associated with big avalanches with initial snowpack failure at or near the ground.

The bond formation rate increases rapidly as the snow temperature increases. Field observations clearly show this effect on avalanche formation. For example, dry snow that falls at warm temperatures commonly stabilizes in place without avalanches developing due to rapid bond formation in new snow. However, this is only one aspect of temperature dependence that is of interest to avalanche forecasters. In general, rounded forms that grow at slow rates are small and tend to pack closer together; therefore, they have more bonds per unit volume and greater strength. Faceting in the presence of high temperature gradients can produce larger grains with fewer bonds per unit volume. In addition, these angular grains do not pack as closely as rounded grains. The overall effect of this is to produce weaker snow.

In addition to the shape and size of individual grains,

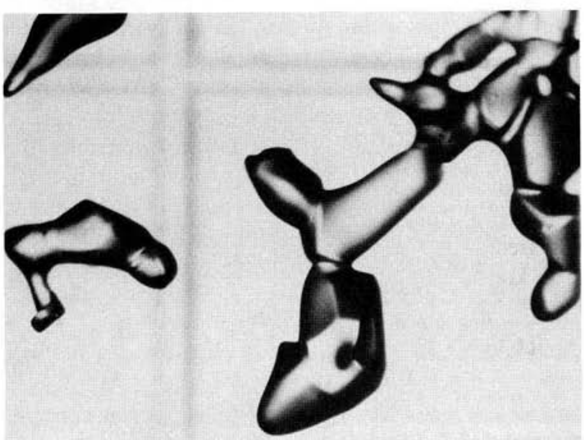


Figure 3.26. Close-up of bonds among grains. Grid size 2 mm. (Photo by R. Perla)

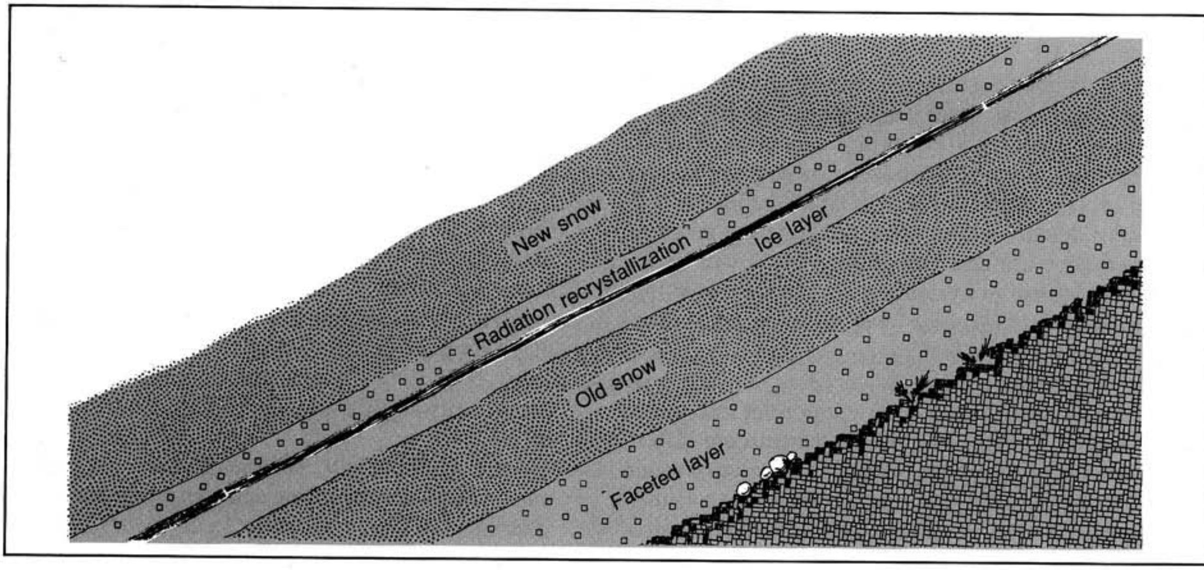


Figure 3.27. Snowpack stratigraphy from a continental climate with the radiation recrystallized layer above an ice layer (see Chapter 2) and the weak faceted layer above the ground.

bond formation between layers of snow also depends on the geometry of adjacent grains between the layers. For example, in general, new snow crystals do not bond well to an ice layer. Also, when the degree of riming changes during a storm, unrimed crystals can experience difficulty bonding to their more heavily rimed neighbors or to each other. Crystals broken by wind packing are usually small and they pack well into a cohesive strong slab. If they bond poorly to the surface below, a condition promoting avalanche formation may be imminent.

The temperature, temperature gradient, grain geometry, and pore space configuration all combine to produce the complicated patterns of bond formation and layer strength observed in dry snow. Avalanche forecasters should be aware of the general concepts with respect to bond formation. However, since it is virtually impossible to understand precisely what is happening at a given instant in time, it is better to use direct methods of assessing snow stability (see Chapter 6) conditioned by a knowledge of the physics of bond formation.

GROWTH OF CRYSTALS AROUND CRUSTS IN DRY SNOW

The influence of a crust can significantly alter the conditions for transport and deposition of water vapor in a dry snowpack. Faceted crystals sometimes form above crusts to produce a future, serious avalanche situation when buried by subsequent snowfall. In fact, weak bonding of snow above crusts is the most important feature of crusts with respect to avalanche formation. Radiation recrystallization (Figure 3.27) provides an example. However, the presence of crusts, since they are relatively impermeable to vapor transport, can also allow faceted snow to grow immediately *below* a hard layer even when the surrounding crystals are well rounded. Below a crust, it is believed that the extra water vapor present due to blockage causes the excess vapor density (supersaturation) near crystal surfaces to increase, allowing facets to form when they might otherwise not form (see the right side of Figure 3.5).

BOND MELTING AND FORMATION IN WET SNOW

The heat flow between particles of different sizes (radius) causes metamorphism and bond melting or formation in wet snow. Efficient heat transfer through the water is the principal element that produces the observed rapid growth (large particles) and decay (small particles and bonds) of ice grains in saturated snow.

When grains touch in saturated snow, the equilibrium *melting* temperature at contact points decreases as the *pressure* between grains increases. (The pressure that forces the grains together is due to the weight of the snow above.) The melting temperature also decreases as the *area* of contact decreases. These effects combine to promote melting at grain contacts and densification (since the distance between particle centers decreases). When melting takes place at grain contact points, the contact area increases. The loss of bond strength at high saturations due to melting of contact points (a result of combined pressure and curvature effects) can be an important element in avalanches that initiate over impermeable layers such as ice layers or the ground where the water saturation is high.

When the water content is very low, surface processes are thought to play a very important role in the mass transfer. In this case, minute temperature differences (caused by curvature) between the region of water tension (bond) and the grain surface determine whether heat flow is toward the area affected by surface tension (in which case melting takes place) or away from the bond (resulting in freezing and true bond formation). If bonds form in this way, they are, of course, much stronger than "bonds" produced by the surface tension of water.

At low water contents, clusters of grains form near the snowpack surface where melting and freezing cycles can occur due to day warming and night cooling, particularly in late winter or early spring. After night cooling this combination produces very strong crusts composed of frozen grain clusters that lose almost all their strength after midday heating. The mass transfer over the crystal surfaces during melt causes the average grain size to increase, and the small grains and bonds disappear. The end result after a number of day–night cycles is clusters of large grains. This process is called *melt–freeze metamorphism.* Melt–freeze metamorphism builds crusts that (if buried) can serve as future sliding layers for avalanches. During the melt phase, wet loose snow may be formed at the surface to initiate wet surface avalanches (see Chapter 4).

References

Bader, H., et al. 1939. Snow and its metamorphism (Der Schnee und seine Metamorphose). Translated January 1954. Snow, Ice and Permafrost Research Establishment, U.S. Army, 313 pp.

Breyfogle, Steven R. 1986. Growth characteristics of hoar frost with respect to avalanche occurrence. *International Snow Science Workshop Proceedings*, Lake Tahoe, California, October 22–25, 1986. Homewood, CA: ISSW Committee, pp. 216–222.

Colbeck, S. C. 1975. Grain and bond growth in wet snow. IAHS Publication No. 114, pp. 51–61.

Colbeck, S. C. 1980. Thermodynamics of snow metamorphism due to variations in curvature. *Journal of Glaciology* 26(94): 291–301.

Colbeck, S. C. 1983. Theory of metamorphism of dry snow. *Journal of Geophysical Research* 88(19): 5475–5482.

Colbeck, S. C. 1983. Ice crystal morphology and growth rates at low supersaturation and high temperature. *Journal of Applied Physics* 54(5): 2677–2682.

Colbeck, S. C. 1985. Temperature dependence of the equilibrium form of ice. *Journal of Crystal Growth* 72: 726–732.

Colbeck, S. C., E. Akitaya, R. Armstrong, H. Gubler, J. Lafeuille, K. Lied, D. McClung, and E. Morris. 1990. The international classification for seasonal snow on the ground. International Commission on Snow and Ice, 23 pp. (Available from World Data Center on Glaciology, University of Colorado, Boulder, CO.)

Fletcher, N. H. 1970. *The Chemical Physics of Ice.* Cambridge, UK: Cambridge University Press, 271 pp.

Hobbs, P. V. 1974. *Ice Physics*. Oxford: Clarendon Press, 837 pp.

Knight, Charles, and Nancy Knight. 1973. Snow crystals. *Scientific American* 228(1): 100–107.

LaChapelle, E. R. 1969. *Field Guide to Snow Crystals.* Seattle, WA: University of Washington Press, 101 pp.

Lang, R. M., B. R. Lee, and R. L. Brown. 1984. Observations on the growth process and strength characteristics of depth hoar. *International Snow Science Workshop Proceedings,* Aspen, Colorado, October 24–27, 1984, pp.188–195.

Magono, Choji, and Chung Woo Lee. November 1966. Meteorological classification of natural snow crystals. *Journal of the Faculty of Science,* Hokkaido University, Sapporo, Japan, Series VII (Geophysics), II(4): 321–335.