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THE MOUNTAIN SNOWPACK

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Snow is one of the most variable substances found in nature and consequently one of the most complex. A high degree of compressibility and thermodynamic instability are two of its dominant characteristics. Examples of these and other characteristics are available from a variety of field and laboratory measurements. New snow crystals begin to change as soon as they collect on the ground or the old snow surface. The type and rate of change depends primarily on temperature conditions. The development of several snowpack types will be presented with emphasis on their respective layered structure. How the layers are formed and why they possess their particular physical properties will be discussed. Layering is the combined effect of the particular meteorological conditions at the time of new snow accumulation and changes (metamorphism) which occur within the snow cover due to variation in overburden pressure and temperature. Processes involved in three distinct types of metamorphism will be described. The concept of the layered structure as the basis for slab avalanche formation will be introduced.

Mechanically, snow exhibits visco-elastic properties. The viscous qualities of snow allow it to deform slowly, in some cases, without fracture. The elastic properties of snow allow energy to be stored (as in a stretched rubber band) and this sets the stage for the brittle type fracture (as a pane of glass would break) associated with slab avalanche release. While many materials have well defined failure criteria (e.g., a definite fracture stress can be determined), snow does not. How snow reacts to the application of stress is determined by its complex and interrelated physical properties as well as the rate of stress application. When snow lies on a slope the relationship between strength and stress becomes important. In general, the stress condition is simply related to the mass of snow on the slope and the slope angle, but strength is controlled by the complex properties of numerous individual layers.

SNOW METAMORPHISM AND INTERGRANULAR BONDS

The International Classification for Seasonal Snow on the Ground has been developed by the International Commission on Snow and Ice (IAHS) and its use is generally accepted. The features used to classify snow include: density, temperature, grain size, hardness (strength), and free water content which in one way or another can be determined in a *reasonably quantitative fashion; and, finally, grain shape or type*, which must be determined by means of an accepted morphological classification scheme. While the international classification of grain types is generally understood and accepted by experienced field workers, the physical processes responsible for the development of these respective stages of snow metamorphism are often unclear or confusing. In the following discussion the intent is to combine the international classification of grain type or morphology with a description of the possible processes which may lead from one identifiable form to another.

Snow is a porous, permeable aggregate of ice grains with its pores filled with air and water vapor and sometimes, liquid water. Many of the unique properties of snow are due to the fact that, unlike most other solid materials found in nature, the ice contained in snow is normally encountered at temperatures close to its melting point. Because of this fact, the water molecules making up the ice particles are relatively free to move around on the ice surface or to actually leave the ice structure and sublimate into the vapor phase. As a result of this mobility, water molecules will eventually be redistributed within a given volume of snow such that the original shapes disappear and distinct new forms are created. This process is called snow metamorphism and it begins as soon as the snow reaches the ground or the old snow surface. The term "metamorphism" means the same as it does in geology, the changes in texture which occur as a result of temperature and pressure.

The rate of metamorphism depends primarily on temperature; the closer the temperature is to 0 degrees C, the faster the change. The process behind these changes involves sublimation at certain locations on the snow grains and deposition at others. Within a given subfreezing snow layer, once the new snow crystals have lost their original shape, the individual grains will tend toward one of two general forms. They may become smooth, rounded grains with average diameters of about 1.0 mm or they may become coarse, angular, or faceted grains with average diameters between 2 and 6 mm or larger. Which path of metamorphism a given snow layer takes is of importance in avalanche hazard evaluation because the rounded, smooth grains tend to create a snow texture which gains strength with time, while the coarse, angular grains show little or no gain in strength with time. What then are the conditions which determine the specific metamorphism of a given layer?

Let us first have a look at how the original shape of the new snow crystal is lost. In order to describe this process in detail we will begin with the well accepted assumption that the air within the pore space of the snow is nearly saturated with water vapor with respect to ice at the prevailing temperature. This means that the same number of water molecules are leaving the ice surface as are arriving back at the ice surface (100% relative humidity). When viewed on a large scale, the system is in balance. This is not the case at the microscale. This is because the total number of water vapor molecules (vapor pressure) which can be supported in a saturated condition just above the ice surface is influenced by the shape of the surface. More molecules can be supported above a surface with a high positive curvature, (convexities, sharp corners, small grains) than over a flat surface, and more molecules can be supported over a flat surface than over a surface with a high negative curvature, (concavities and grain contacts). This difference in vapor concentration causes a net flow of molecules from the locations of higher concentration to those of lower concentration resulting in the loss of sharp angles on the crystal faces and the gradual development of a more rounded or isometric shape. However, it is currently believed that this process is of importance only when radii of curvature (positive or negative) are very small, e.g., 1.0 to 100.0 microns.[1] Such radii would be typical of the angles on the arms and branches of new unrimed stellar crystals.

About the time the original shapes of the new snow crystals are no longer recognizable and the sharpest angles (smallest radii of curvature) have been lost, the influence of surface curvature begins to diminish. At this point the specific course of metamorphism to be followed by a given snow layer will be primarily determined by the temperature gradient. A *temperature gradient* is the variation in temperature with distance measured up or down in the snow cover. In a natural snow cover, the temperature gradient usually varies from warmer near the ground to colder near the snow-air interface. As was shown in the section on meteorology, warm air can hold more water vapor than cooler air. Therefore, the air which fills the pore space within snow cover near the ground contains more water vapor than is contained within the colder layers above. In terms of conditions at the surfaces of the individual ice grains, this means that the concentration of vapor is greater adjacent to

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[1] (0.001 to 0.1 mm)

the grains in the warmer layers than it is adjacent to the cooler grains. This difference in vapor concentration results in a vapor pressure gradient. Because this vapor pressure gradient is produced by the temperature gradient, the greater the temperature difference the greater the vapor pressure gradient. In the presence of a strong temperature gradient (10 to 20 degrees C/m or more), a relatively rapid transport of water vapor occurs due to this imbalance in water vapor concentration. As the water vapor moves from warmer to colder locations in the snow cover, it comes in contact with colder grain surfaces resulting in a condition of high supersaturation. Consequently, some of the vapor is deposited as solid ice on the face of the colder grain. A series of sublimation and deposition processes between grains takes place with the relatively warmer ice surfaces losing mass to the colder surfaces. This is generally understood to occur on a rather small scale, with the transfer taking place from one adjacent grain to another, i.e., the upper surface of one grain is the location with the slightly higher temperature and the lower surface of the grain just above it is the cooler surface. In the presence of a strong temperature gradient, grain growth occurs by the evolution of visible steps across the crystalline faces. This process produces specific shapes characterized by facets, steps, and eventually hollow cup crystals. Grain growth may be quite rapid and the smaller grains disappear creating fewer but larger grains. This process used to be referred to as "Temperature-Gradient (TG) Metamorphism," because the bulk of the vapor transfer results from temperature gradients in the snow cover. However, this term is somewhat misleading since all snow metamorphism outside the laboratory takes place in the presence of at least a small temperature gradient. Thus, we now simply refer to this process by the crystal forms associated with it: "**Faceting**". A sustained strong temperature gradient (20-40 degrees C/m) in new low density snow can create grains 3 to 6 mm in diameter in approximately 10 to 20 days.

The rate of crystal growth is, however, not only controlled by temperature gradient, but also by the average temperature of the gradient. The closer the temperature is to 0 degrees C, the more water vapor that is available for transport within the pore space. In fact, the rate of crystal growth is actually controlled by the vapor pressure gradient, which is itself determined by the overall temperature conditions. The relationship between temperature and vapor pressure is non-linear and is shown in Figure 1. Steep temperature gradients alone do not always provide adequate vapor pressure gradients. For example, a temperature gradient of 10 degrees C/m may result in moderately rapid crystal growth if the average temperature of the snow layer is very close to 0 degrees C. At mid-latitude sites, this would commonly occur in those layers at the base of the snow cover near the warm ground.

However, as the average temperature of the snow layer decreases, the temperature gradient must increase to provide the same vapor pressure gradient and associated crystal growth rate. If, for example, the average temperature of the snow layer decreases from near 0 degrees C to -20 degrees C, the temperature gradient must be increased nearly five times in order to provide the same vapor pressure gradient, (See Figure 1).

When the temperature-gradient is less than about 10 degrees C/m, the resulting grains have a shape totally different from those produced by faceting. Under these conditions, water vapor is still moving from the warmer to the colder portions of the snow cover but at a much slower rate. This causes grain growth to be much slower. In this case, each site on the grain surface is more or less equivalent and the water vapor is deposited uniformly over the surface. The result of this slow growth rate is a grain with smooth surface features and a rounded shape is approached. This is in contrast to the process just described where strong temperature gradients and rapid growth rates produce highly preferential vapor deposition at specific locations on the grain surface, and distinctive coarse, angular grains result. This slow growth process was previously known as "Equitemperature (ET) Metamorphism". However, it does occur in the presence of a weak temperature gradient (in fact, when there is no temperature gradient the process is slowed significantly), so we now refer to this process simply as "**Rounding**". In the presence of a weak temperature gradient, another process known to occur is sintering. ***Sintering*** is the growth of bonds between the grains producing an increase in intergranular strength. When the concavities at the grain contacts are in the form of sharp, negative angles, the sintering process will be driven by differences in curvature, as described above. In this case, molecules migrate from the flat or slightly convex surfaces to the sharp, negative angles of the grain contacts. As the vapor deposits at these locations, the bonds or necks between the grains grow and strength increases. This process will take place in the absence of a temperature gradient, but it is understood to take place at a slightly accelerated rate in the presence of a weak temperature gradient, (less than 10 degrees C/m). In the presence of a strong temperature gradient, the rapid movement of vapor apparently overrides the sintering process and little or no bond growth occurs.

In summary, the average temperature of the snow layer determines the *rate* of metamorphism, being faster the closer the temperature is to 0 degrees C, while the temperature gradient determines the *type* of metamorphism. Strong temperature gradients eventually produce large, faceted, and poorly bonded grains while weak temperature gradients result in relatively small, smooth, rounded grains with some degree of intergranular bonding.

The processes described above may occur as long as the temperature of the snow remains below 0 degrees C. If the snow temperature reaches 0 degrees C, wet snow or "**Melt-Freeze (MF) Metamorphism**" will result. A thorough description of the warming of the snow cover to 0 degrees C and the subsequent production of liquid water (free water) is presented in other parts of this notebook dealing specifically with wet snow, so only a brief summary and a few additional details will be presented here. Wet snow metamorphism is a more complex process due to the fact that it involves a three-phase system: ice, vapor, and now, liquid water. Nevertheless, the same physical principles drive the metamorphic processes and again, due to curvature differences, large grains will grow at the expense of smaller grains. The continuation of a diurnal melt-freeze cycle tends to diminish the number of grains while increasing the average size of the grains. This results from the fact that smaller grains have a slightly lower melting temperature and, therefore, will melt first. However, at night the liquid water resulting from these melted grains refreezes onto adjacent ice grains. After several melt-freeze cycles, clusters of large, coarse grains develop, and these are called melt-freeze grains.

Rain can be a significant source of free water, but it is not often important in producing melt. In winter, at a high altitude site, rain is rarely more than a few degrees above freezing. Under these conditions, rain is not a significant heat source with respect to melt. For example, with rainfall at +4 degrees C, approximately 12 hours of rain would be required to melt one centimeter of snow.[2] Rain falling on a subfreezing snow cover will refreeze and the associated release of latent heat will warm the surface layers. When free water, produced by rain and/or melting, percolates downward through 0 degrees C surface layers and encounters colder layers, the free water will refreeze, again releasing latent heat. This process allows that portion of the snow cover still below 0 degrees C to be warmed in a very efficient manner. For each

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[2] At 4 degrees C each gram of rain would contain 4 calories of heat. Eighty calories are required to melt one gram of ice. Therefore, 20 grams of 4 degrees rain are required to melt one gram of ice. One gram of ice is equal to 1.1 cm^3 of ice or approximately 4 cm^3 of snow with a density of 250 kg/m^3. With a moderate rainfall rate of 3 mm/hour, which is equal to 0.3 grams of water per hour, it would have to rain for more than 66 hours to melt the 4 cm^3 of snow, (20 grams - 0.3 = 66.7). In summary, it is safe to say that more than 12 hours of steady rain at 4 degrees C would be required to melt even 1 cm of snow with a density of 250 kg/m^3.

cm^3 of water that refreezes, 80 calories are released. This amount of heat has the capacity to increase the temperature of more than 400 cm^3 of snow by 1.0 degrees C.[3] In addition, the glaze or crust formed by rain falling on cold snow will decrease the albedo of the snow, thus increasing the amount of solar radiation which can be absorbed at the snow surface.

With respect to avalanche release, the primary significance of rain on snow is the addition of weight to a slope. A moderate to heavy rainfall often provides additional mass to a slope much more rapidly than typical snowfall rates. In addition, the added weight of the rain does not provide the additional tensile strength that would accompany a layer of new snow.

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[3] When one gram of water refreezes, 80 calories of latent heat are released. The specific heat of ice is 0.5 calorie/degree C which means that only 0.5 calorie is required to raise the temperature of one gram of ice 1 degree C. Therefore, the refreezing of one gram of water (which is equivalent to 1 cm^3 of water) releases enough heat to raise the temperature of 160 grams of ice 1 degree C, (80 calories - 0.5 calories). One hundred sixty grams of ice are equal to 174 cm^3 of solid ice in terms of snow with a density of 400 kg/m^3 would be 435 cm^3. Therefore, the refreezing on only 1 gram of water would have the capacity to warm a volume of snow about 7.6 cm on a side (=435 cm^3) by 1 degree C.

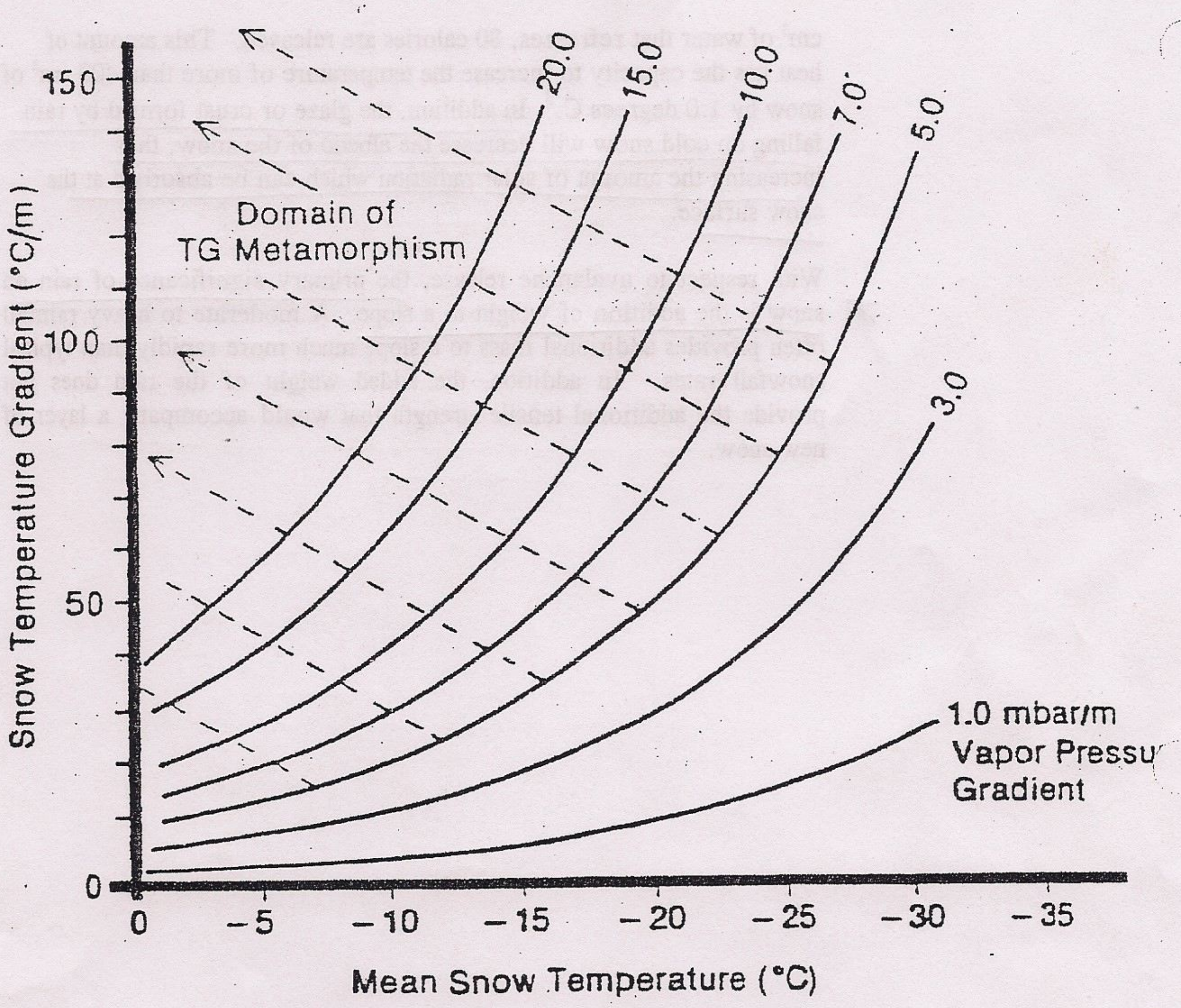

Figure 1. *The relationship of vapor pressure gradient (in millibars per meter) to temperature gradient and average temperature within the pore space of a snowcover. The value 5.0 mbar/m is considered by this author to be the lower limit for rapid TG metamorphism in low density new snow.*