WET SLAB INSTABILITY[1]

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Abstract.--Wet slab avalanches often occur when liquid water reduces the shear strength between adjacent snow layers. Identification of layer interfaces where water may accumulate is an important forecasting tool. This paper reviews recent literature and current work on water movement in snow with respect to evaluations of snow stability.

INTRODUCTION

As commonly perceived in the continental United States, wet snow avalanches constitute much less of a problem than do dry snow avalanches. Nevertheless, they can be very important in some regions and occur in almost all mountain areas, and occasionally cause severe damage.

The occurrence of wet snow avalanches largely depends on the snow climate of an area. In maritime snow climates or areas of "warm" snowpacks, such as the west coasts of Japan and North America, a large proportion of wet snow avalanches occur during winter, as a result of wet snowfall, rain, and midwinter snowmelt. In continental snow climates or areas of "cold" snowpacks, wet snow avalanches occur almost exclusively in spring. The warm snowpacks tend to be relatively stable in spring during active melt.

A primary difference between dry and wet avalanches is that dry snow usually fails due to an increase in shear stress while wet slides usually occur because of a decrease in shear strength. If this critical decrease in strength occurs along an interface between cohesive snow layers or at ground level, a slab avalanche will be released. Alternatively, if there is a general loss of strength between all of the snow grains in one or more of the upper layers, the avalanche will be the wet, loose type, usually beginning at a point near the snow surface. Slush flows represent an extreme of wet avalanche conditions in which large quantities of water moving through the snow cause the entire mass to flow. This paper reviews some of the literature and recent work in the Sierra Nevada concerning liquid water in snow with respect to snow stability. It concentrates on the wet slab type, although much of the material also applies to loose snow avalanches.

METAMORPHISM OF WET SNOW

Instability of wet snow begins with the introduction of liquid water into dry snow. When a small amount of water is added to dry snow, the rates of growth and rounding of the snow grains dramatically increase (Wakahama 1968). As liquid water content increases, the rate of metamorphism increases slowly until a critical level is reached where the growth rate again increases markedly and a shift in processes occurs (Colbeck 1973b). This theory is useful for description and is widely, though not universally, accepted. At water contents below this critical level, air is continuous throughout the pore space and water is in isolated cells. At higher water contents, water is continuous throughout the pore space and air occurs only in isolated bubbles.

The terms "pendular" (low water content) and "funicular" (high water content) have been applied from soil physics to distinguish between these two conditions (Colbeck 1973b). The transition between the two regimes was theorized to occur when about 14 percent of the pore volume was filled with water (Colbeck 1973b) and experimentally observed in a range of 11 to 15 percent of the pore volume (Denoth 1980). In terms of the more commonly used volumetric water content, the pendular-funicular transition occurs at about 7 percent (Colbeck 1982). The distinction between these two regimes should be recognized, because the rates and processes of snow metamorphism and the consequent water flow rates and mechanical strength differ considerably between them.

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Pendular regime

The pendular regime is characterized by air in continuous paths throughout the pore space which limits the area available for heat flow and by isolated regions of water under tension (Colbeck 1973b). The capillary pressure is quite high and greatly influences the processes of metamorphism. The presence of water results in an increase in grain growth (Wakahama 1968) and clustering together of the grains (Colbeck 1979). Both the growth and clustering of grains occur as the snow approaches a more stable state (Colbeck 1982).

Snow is naturally a mixture of different sized grains, each existing at a slightly different temperature (on the order of thousandths of a degree Centigrade). Larger grains have slightly higher equilibrium temperatures than do smaller grains. This temperature difference leads to heat flow from the larger grains to the smaller grains and results in melting of small grains while larger grains increase in mass. In this pendular regime, the contiguous air in the pore space severely slows down these heat and mass flows. Mass transfer appears to occur by diffusion over the surfaces of the grains (Colbeck 1982) in a quasi-liquid layer (Ushakova and Troshkina 1975). The capillary pressure of snow in the pendular regime is great enough to influence the temperature of the snow mass and, thus, the temperature differences between grains. As the water content increases, the capillary pressure decreases, and the temperature differences between grains become more effective, increasing the rate of metamorphism (Colbeck 1973b). As the water content decreases, the capillary pressure increases, and the intergrain attraction is increased resulting in grain clustering and high mechanical strength (Colbeck 1973b, 1982).

Funicular Regime

The funicular regime is characterized by contiguous water surrounding the snow grains and, thus, the absence of a solid-gas interface. Experiments in Japan demonstrated that large grains grew round and even larger at the expense of smaller grains, which decreased in size and ultimately disappeared, reducing the total number of grains (Wakahama 1968, 1975). The growth rate decreased from 0.02 mm/hr to zero after 3 days. The number of intergrain bonds decreased with time.

The rate of grain growth is much higher during active melt from external energy input than it is under equilibrium conditions. In the funicular regime, grains grow rapidly in response to the difference in temperature between grains of different sizes (Colbeck 1973b). This growth occurs much faster than in the pendular case owing to the relatively unrestricted heat and mass flow between the large and small grains through the liquid (Colbeck 1982). Grain growth is ultimately limited by the disappearance of the small grains. The end product of metamorphism in the funicular regime is a snow matrix composed of larger grains with fewer grain-to-grain contacts per unit volume than in the initial condition (Wakahama 1968). Mechanical strength of the snow matrix is further reduced by melting at the intergrain contacts resulting from compressive stress between the grains and pressure from the overlying snow (Colbeck 1973b).

MECHANICAL STRENGTH OF WET SNOW

Understanding of wet snow mechanics is not as well developed as for dry snow (Salm 1982). Long before the theories of wet snow metamorphism were formulated, the cohesiveness of snow was observed to decrease as liquid water increased (Seligman 1936). Likewise, the angle of repose of snow is known to drop drastically (Perla 1980), and creep and glide velocities to increase (Schaerer 1981) with increasing liquid water content. High avalanche activity has been directly related to measured liquid water contents above 7 percent by volume (Ambach and Howorka 1965). As an index of the decreasing strength of snow with increasing liquid water, the Kinosita hardness in a snowpack was found to decrease by two orders of magnitude as liquid water increased up to 20-25 percent (Wakahama 1975). In summarizing snow mechanics, Salm (1982) stated, "There is still a long way to go until...the release of a wet snow avalanche can be fully understood."

MOVEMENT OF WATER THROUGH SNOW

In the past decade, a reasonable understanding of water flow through snow has been developed (Colbeck 1978, Wankiewicz 1979, Jordan 1983a). The gravity theory of water percolation through snow was evolved by Colbeck in a series of papers beginning in 1972. In its simplest form, the theory explains that water flows through snow primarily in response to gravity and is an application of Darcy's law: $V = -K\ (d\psi/dz + 1)$, where the flux V in meters/second is proportional to the sum of the capillary pressure gradient ($d\psi/dz$) and the gravitational pressure gradient (1 m/m). The constant of proportionality (K) is the unsaturated hydraulic conductivity. Gravity drainage is predominant during steady or decreasing flow, and the capillary pressure gradient may effectively be ignored except at some interfaces between snow layers (Colbeck 1974). Later experiments demonstrated that a negligible capillary pressure gradient exists throughout the snowpack as a whole, but that significant pressure gradients may occur at major pressure or textural discontinuities (Wankiewicz 1978).

Modeling of water flow through snow on the basis of the gravity flow theory has been extended to include the effects of ice layers and flow channels (Colbeck 1979, Marsh 1982), increasing flow (Jordan 1983b), and refreezing and water retention (Bengtsson 1982). The two most complex problems associated with routing water through a snowpack involve flow interactions with layers in the snowpack and concentrations of flow in distinct channels. Both of these phenomena were described in the Alps half a century ago (Seligman 1936).

Ice Layers

High density layers, often referred to as "ice layers", have been observed to restrict the flow of water in Sierra Nevada snowpacks (Gerdel 1948, 1954; U.S. Army Corps of Engineers 1956). These layers were believed to increase water storage in the snowpack for only a short time before they disintegrated. If inclined, the layers could route water downslope (Gerdel 1954, Langham 1973a). While water has been observed to flow over an impeding layer for distances exceeding 2 meters, it is not possible to estimate any sort of an average flow distance for various conditions. The thickness and permeability of ice layers appear to vary both spatially and temporally (Langham 1973, 1975; Marsh 1982). Their permeability depends on intergranular veins, which expand and contract with temperature. Ice layers may form through several mechanisms: repetition of melt-freeze cycles of the snow surface and subsequent burial (Berg 1982); freezing of liquid water above a snow layer interface (Colbeck 1982, Marsh 1982); or addition of freezing rain to the snow surface and its subsequent burial (Langham 1973).

Water can also be impounded at layer interfaces where texture changes abruptly between fine-grain snow above coarse-grain snow (Wakahama 1968, 1975; Colbeck 1973b). A volumetric liquid water content of up to 30 percent has been measured above such interfaces (Wakahama 1975). As stated earlier, capillary effects become important at textural discontinuities. Since fine-grain snow contains smaller pores than does coarse-grain snow, capillary attraction will be greater in the fine-grain snow (that is, water will be held more tightly in the smaller pores). When fine-grain snow overlies coarse-grain snow, water will accumulate in the upper layer until the pressure difference between the layers is relieved and water can flow into the lower layer (Wankiewicz 1979).

In the extreme case where snow overlies an air space, water accumulates in the snow until the water pressure at the interface equals atmospheric pressure, at which point water begins to drip. In these situations, a saturated layer some 2 cm (Colbeck 1974) to 3 cm (Wankiewicz 1976) thick will form. If the air space is replaced by a coarse-grain snow layer, water will still accumulate but to a lesser extent, and water will begin to flow at a pressure somewhat less than atmospheric, depending on the pore size. In theory, a coarse-grain snow layer overlying a fine-grain layer would accelerate flow (Wankiewicz 1979). However, all snow layer interfaces, regardless of type, restricted flow between the layers in the High Arctic (Marsh 1982).

When a saturated snow layer above an ice layer or texture interface is inclined, water will flow downslope over the impeding boundary (Gerdel 1954, Colbeck 1973a, Wankiewicz 1979). The saturated layer is a prime example of the funicular regime in which grain growth will be rapid. Because it has large grains with continuous water paths between them, the layer has high permeability and enhances downslope flow (Colbeck 1973b). Such flow increases rapidly as the slope increases from zero to 10° and then increases relatively slowly for greater slopes (Colbeck 1973a). The thickness of the saturated layer will increase downslope as water accumulates both from vertical drainage and downslope flow. In the presence of the saturated layer, the permeability of the impeding layer below will increase. At some point downslope, the water will begin to flow into the lower layer through a vertical drain (Colbeck 1973a, Wankiewicz 1979).

Flow channels

Vertical drainage channels are composed of snow of above average grain size and permeability (Gerdel 1948, Wakahama 1968). Concentrations of flow occur below textural discontinuities in the snowpack. When water crosses an interface, it does so at isolated places, whether the endpoint is an intergranular vein traversing an ice layer or the area where pressure between layers of different textures was first equalized. Such zones where water enters a layer at a point or in a localized area are often called "flow fingers". These channels become the preferred pathways for additional water flow as grain size and permeability increase, thus developing a positive feedback mechanism. The presence of water leads to larger grains and greater permeability, which in turn allow more water to flow.

Channels that conduct water appear to become better developed and are more important in warm snowpack areas than in cold snowpack areas. In subfreezing snow, the advance of water in flow fingers is severely limited by the cold snow which freezes the percolating water while increasing the latent heat release (Marsh 1982). Meanwhile, a general wetting front moves downward only a few centimeters behind the advance of the fingers. In addition to the subfreezing situation, the isothermal case was also modeled by Marsh (1982). Because little internal freezing occurs in the 0°C snowpack, flow fingers develop rapidly through the profile. According to this model, water in flow fingers reaches the base of an isothermal snowpack 4 days before the general wetting front as opposed to only 12 hours for a cold snowpack in the High Arctic. With this long period in which the grains may grow, the flow fingers in an isothermal snowpack will have a much higher permeability than the surrounding snow and thus should continue to conduct most of the water flowing through the snowpack. In cold snowpacks, this large difference in grain size between the fingers and the snow matrix does not have a chance to develop (Marsh 1982).

A special type of channel has recently been observed in Sierra Nevada snowpacks (Kattelmann, in press): open channels devoid of snow which appear to be capable of conducting large quantities of water. They are analogous to the well known soil macropores. The open channels in snow are found in the basal layer of the snowpack and appear to require saturated conditions for formation. Once formed, these snow macropores should allow rapid drainage of the lower part of the snowpack.

Water Retention

The amount of water stored in the snowpack can also influence avalanche activity. Water entering a cold, dry snowpack will be frozen, thus increasing snowpack mass. After the release of latent heat warms the snow to 0°C, additional water is stored in pore spaces as capillary water in addition to hygroscopic water on grain surfaces. The liquid water held against gravity is often called the "irreducible water content" (Colbeck 1972). This irreducibility should be considered only at a particular metamorphic stage. As grains enlarge, so do pore spaces, and the capillary retention decreases. Characteristic values of water retention in snow have not been determined; available data include a wide range of values (Wankiewicz 1979).

Initial results from a study at the Central Sierra Snow Laboratory of the USDA-Forest Service at Soda Springs, California, indicate that overall water retention in maritime snowpacks is fundamentally related to water distribution in snow (as traced by minute quantities of a dye). While areas that conduct liquid water may retain in excess of 10 percent liquid water by volume one day after wetting, such areas occupy only 10-20 percent of the volume of the snowpack. Even minor slope angles appear to further constrict the volume wetted by flowing water and thus the water holding capacity, as well. No consistent relationship between water retention and grain size, density, or age has yet been found. If the snow has been wetted previously, it does not appear to be capable of retaining additional water. Observation of snow profiles after multiple dye applications showed that water flows in the previously established paths. Measured water holding capacity one day after wetting ranged from 0 to 7 percent by volume. The reproducibility of these measurements, based on changes in density corrected for changes in volume, has been poor, indicating a high degree of variability in the snow property itself, as suggested by Wankiewicz (1979).

WATER AND SNOW STABILITY

Current understanding of the interactions between snow and liquid water has a variety of implications for evaluating the stability of wet snow. In general, mechanical strength decreases as liquid water increases. In the presence of water, snow has fewer small grains, fewer points of contact, fewer bonds, and relatively low strength at these bonds. While a very low water content binds snow together by capillary forces, this attraction decreases quickly at higher water contents. A major decrease in strength occurs when the water in the pores becomes continuous at about 7 percent liquid water by volume (Colbeck 1982). While it is probably not worthwhile in most areas, liquid water measurements could be incorporated in an avalanche forecasting scheme. If so, the dilution technique (Perla and LaChapelle 1984, Davis and Dozier, in press) is probably the most promising method of measurement at the present. An electromagnetic technique that is practical for remote use is needed. Nevertheless, an assessment of where water might concentrate within the snowpack might be of much greater value. A set of mathematical models exists to analyze the flow of water through snow. While such tools have been applied to avalanche forecasting (e.g., Carroll 1978), they probably will not be widely used in this field. Nevertheless, their development has provided many applicable principles.

Water will accumulate above high density ("ice") layers and interfaces with even a slight discontinuity in texture. Such layers are rarely completely impermeable or of low permeability for very long. However, significant grain growth can occur in just a few hours under the funicular regime (Raymond and Tusima 1979). When such metamorphism occurs at a distinct layer over much of an area, serious instability exists. The accumulated water can also act to lubricate the zone of shear, so that in the event of slab failure, the potential for downslope motion is greatly enhanced (Seligman 1936). If water is flowing downslope over the layer of restricted permeability, additional force may be imposed upon the overlying slab (Moskalev 1966). The instability may be relieved by any of the following: avalanches, freezing of the ponded water (Seligman 1936) and/or the slab itself, deterioration of the ice layer (Langham 1974) and subsequent drainage, or sufficient melting of the overlying slab to reduce the stress at the shear plane.

Almost any textural interface should be considered as a possible impeding layer and potential sliding surface. Thick ice layers are certainly the most obvious stratigraphic feature, but interfaces of fine-grain snow above coarse-grain snow may be just as likely to cause ponding of water. Indeed, layers so thin that they could be easily overlooked in describing the stratigraphy may have a major influence on routing of water in the snowpack (Gerdel 1948). Daily applications of dye to the snowpack before and during the period of concern can illustrate potential problem layers upon excavation. Applying dye as a powder to the surface is preferable to spraying dyed water because unnaturally high rates of water application will result in unstable flow patterns that may not resemble natural water distribution. While the dye will accelerate snowmelt, the meltwater input is still at a reasonable level. The use of fluorescent dyes allows very small quantities of the tracer to be used. Alternatively, dyed crystals may be introduced at some level below the surface (Langham 1971).

Snowpacks on ski runs tend to have a poorly defined stratigraphy due to physical disturbance of the snow. Thus, water has little opportunity to accumulate at particular layers. The few problem layers that may exist, such as at the base or at interfaces formed during a long interstorm period, are more easily recognized than are potential sliding surfaces in an undisturbed snowpack.

Stratigraphic profiles of the snowpack near the Central Sierra Snow Laboratory from the winters of 1983 and 1984 have revealed a very complex structure of snowpacks exposed to midwinter water percolation. While major structures such as buried surface crusts from long interstorm periods are identifiable in several profiles, many interfaces and ice layers appear to be limited to a small area. For example, several series of thin ice layers that were quite distinct in one profile, could not be found in another snowpit only a few meters away. Such observations imply that water movement through a maritime region snowpack creates a highly irregular structure that would help keep the snowpack in place by avoiding subsequent concentrations of water in widespread layers. Release of water to the ground surface at isolated points may also be a stabilizing influence by avoiding the widespread lubrication of the snow-ground interface, although this effect is difficult to assess.

The presence of water-conducting channels within the snowpack generally adds to stability by leaving much of the snow volume relatively dry and, therefore, stronger. The presence of dry zones within an arctic snowpack has been observed to provide enough strength to an otherwise unstable snow mass to prevent artificial release (Hamre and Stethem 1980). Flow channels or drains also increase stability by routing water away from layer interfaces and out of the snowpack. Snow macropores, where they exist, allow rapid drainage of the basal snow layer which is a common bed surface. The presence of channels that allow good drainage in warm Sierra Nevada snowpacks helps to explain their springtime stability. Such channels are not as well developed in cold snowpacks (Marsh 1982), and perhaps consequently spring avalanche cycles in the Rocky Mountains are more common than in the Sierra.

As mentioned previously, water holding capacity is related to water distribution and therefore to channeling within the snowpack. Continental snowpacks probably have a greater water retention capacity than do maritime snowpacks, at least because of the greater "cold content" or "heat deficiency" but also probably due to less well developed channels. Snowpacks and snow layers with a high water holding capacity are of greater risk from rain storms because of their ability to retain water, thus increasing the stress. Another destabilizing influence of rain is to induce rapid settling in a surface layer, increasing tension per unit area[3]. Settling rates of up to 25 cm/day have been measured during rain storms at the Central Sierra Snow Laboratory, and even greater rates probably occur over shorter time intervals.

The nature and permeability of the snow-soil interface are other important factors to consider. Relatively little is known about mass and heat flow (Guymon 1979) or adhesion between the soil and snow. Wet snow avalanches are often observed to run on frozen or water-saturated soils, rock, grass, and basal ice layers (Gardner 1983). Surveys of the bare ground in an area of concern will indicate likely problem areas. Such surveys are most informative as the snow cover recedes and immediately after. The location of springs and seeps should be noted, as they can introduce additional water into a potential shear zone. The extent of basal ice layers can be found during winter and early spring snowpit excavations. All such areas of restricted permeability will be areas of water accumulation and potential sliding surfaces.

An Example

A wet slab avalanche cycle in May 1983 in the Lake Tahoe area illustrates several of the characteristics outlined above. After one of the heaviest winters on record, storm activity continued into early May. A storm depositing 30 to 40 cm of snow was followed by 3 days of clear skies and mild (-7 to +6°C) temperatures. A melt-freeze surface crust began to develop. Snow began to fall in the early evening of May 9, and up to 30 cm of snow fell during a 12-hour period. Clear skies and warm temperatures followed. Then, beginning on the afternoon of May 14, avalanches of the last storm layer occurred throughout the area. The slides tended to be small in extent and no incidents were reported.

On May 15, in the early afternoon, I released a small slab of about 100 m^2 area which ran for less than 50 m. The slab was released as the snow on almost level ground below and over 100 m away collapsed under the weight of one skier and the shock propagated up the slope of about 25°. The 20 cm high crown was saturated with liquid water in the lowest 2 to 3 cm. When snow blocks were cut away above the crown, there was virtually no adhesion with the underlying snow. The snow grains of the underlayer were about twice the size of those in the overlying layer. While the difference in grain size was obvious, the upper part of the lower layer could not be considered an ice layer. Excavation of the area above the crown fracture showed no sign of water flowing across the interface but, instead, liquid was ponded above it.

PREDICTION AND CONTROL

Artificial release of avalanches with explosives (Mellor 1973) is not as effective in wet snow as in dry snow (Armstrong 1976) and may require special blasting techniques (e.g., Israelson 1978). Armstrong (1976) suggested, therefore, that precise prediction of wet snow avalanche release is needed. Each geographic region or even a particular slide path may have recognizable patterns of wet snow failure. Special observations, such as monitoring snowpack temperature (Hamre and Stethem 1980) or measuring

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[3] LaChapelle, E.R. 1983. Wet snow. Lecture presented at the National Avalanche School, Reno, Nevada.

water content (Ambach and Howorka 1965), may be of great value in forecasting for a specific area. By relating a sequence of meteorological events to existing snowpack conditions, wet snow avalanches have been forecast with a high degree of accuracy in southeast Alaska (Hutcheon and Lie 1978) and in the San Juan mountains of Colorado (Armstrong 1976). Systematic monitoring of wet snow avalanches and the contributing conditions is needed to provide the basis for forecasting wet avalanches on specific paths. The basic principles of water interaction with snow described here may be applied to a wide range of conditions. Nevertheless, careful observation, synthesis, and judgment are always required for particular stability evaluations.

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