SNOW COMPACTION EFFECTS ON NIGHTTIME FREEZING

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Artificial compaction is among the most drastic of available snow management and engineering techniques. Snow is compacted to improve recreational skiing conditions; to create winter roads, parking lots, and runways; and for use as a building material. Compaction increases both density and hardness of snow, with a variety of consequences.

As part of an assessment of selected hydrologic effects of snow compaction at ski areas, several snowpack properties and processes were studied at compacted and uncompacted sites in the Sierra Nevada (Kattelmann, 1981; 1985). Snowmelt runoff from the compacted areas appeared to occur in a shorter part of each day than runoff from the uncompacted areas. A study in Montana also found more rapidly responsive hydrographs from compacted basins than from uncompacted watersheds (Grady, et al., 1982). One process studied during the 1981 Sierra investigation was nighttime freezing of near-surface snow. The degree of nocturnal freezing affects the timing of meltwater release on the following day and may account in part for observed differences in runoff from compacted and natural snow.

SNOW COMPACTION

Artificial compaction of snow is known to initiate or accelerate two metamorphic processes: densification and hardening (Taylor, 1953; Wuori, 1963). Densification includes rounding of crystals, growth of larger grains at the expense of smaller ones, and reduction of snowpack porosity. Hardening or sintering involves the formation of ice bonds between grains. Both changes occur naturally in the snowpack but are greatly accelerated by artificial compaction. Mechanical disturbance breaks off the small points of new snow crystals, breaks weak pre-existing bonds, and brings the crystals and grains into much closer contact than would occur naturally (Taylor, 1953; Wuori, 1963).

Compaction increases heat flow through dry snow by increasing thermal conductivity and leads to differences in the temperature regime between compacted and natural snow (Colbeck, 1983). Thermal conductivity increases proportionally with the square of snow density (Yen, 1969). Snow and soil temperatures have been found to be colder under compacted rather than uncompacted snow (Neumann and Merriam, 1972; Baiderin, 1980).

NIGHTTIME FREEZING

The formation of a nighttime freeze crust both reduces and delays the quantity of water released from a snowpack subject to some daily energy input (U.S. Army Corps of Engineers, 1956). The near-surface layer of snowpacks that are actively melting during daytime forms a frozen crust when the surface energy balance becomes negative. This net loss of energy occurs on most clear nights as the snowpack emits longwave radiation. In early evening when energy input no longer exceeds energy loss, liquid water in the surface layer freezes and releases latent heat. As additional energy is lost, this layer cools and liquid water in the underlying snow freezes due to diffusion of heat into the colder snow above. The boundary between frozen snow and wet snow moves deeper into the snowpack and the frozen layer becomes colder. This process continues throughout the night as long as energy input is less than energy loss. When the surface energy balance again becomes positive, the crust warms and begins to melt.

Nighttime freeze depth and the internal energy of the snowpack (cold content or heat deficit) can be estimated using simplifications of standard heat flow equations (Bengtsson, 1981; 1982). At best, these equations are indices of heat transfer and storage. Crust depths estimated from these equations compared poorly with measured depths (McGurk, 1985). An estimate of thermal diffusivity, a property that "governs the speed and attenuation of temperature waves propagating into the snow from the surface" (Langham, 1981:298), is required for these calculations. Reported estimates of thermal diffusivity range from 0.0006 to 0.012 cm^2/s (Bengtsson, 1982; Dorsey, 1940; List, 1959; and Yen, 1969). Thermal diffusivity increases with density (Yen, 1969) and changes with the degree of intergrain bonding (Langham, 1981). Therefore, snow compaction should increase thermal diffusivity

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through both densification and sintering. Here, compaction causes an increase in thermal diffusivity, which results in greater nighttime freeze depths and greater heat loss.

Nighttime crust thicknesses are typically less than 15 cm (Bengtsson, 1982; Langham, 1981; Miller, 1955; and U.S. Army Corps of Engineers, 1956) although the snow may freeze as deep as 25 cm on occasion (U.S. Army Corps of Engineers, 1956). When the effect of refreezing (release of latent heat) is considered and a linear temperature distribution is assumed, the total nighttime heat deficit is typically less than 30 joules/cm^2 or 1 mm of melt equivalent (Bengtsson, 1981).

MEASUREMENTS OF FREEZE DEPTH

In order to examine differences in night crust thicknesses between compacted and uncompacted snow, observations were made of freeze depths at two Sierra Nevada sites near Donner Pass on Interstate Highway 80 at 2100 m elevation. Crust thicknesses were measured in April and May 1981, a year of subnormal precipitation in northern California. A level site at the Central Sierra Snow Laboratory (CSSL), a facility of the U.S.D.A. Forest Service's Pacific Southwest Forest and Range Experiment Station, was compacted by tracked vehicles and by foot. The other site was on a north-facing ski run at a nearby ski area. Freeze depths were also measured in adjacent undisturbed snow subject to the same environmental conditions as the corresponding compacted site.

Freeze depth was identified as an obvious discontinuity in hardness (hard frozen snow above this level and soft wet snow below it) along an exposed pit wall. Small snowpits were excavated in early morning to allow examination of the profile. The wall was scraped to find the level where snow grains were easily dislodged and where a finger or pencil could easily be pushed into the snow. The depth from the surface to the base of the frozen layer was then measured by tape and recorded to the nearest centimeter. Generally, five pits were excavated at each site in previously unsampled snow. The maximum and minimum measured depths usually differed by less than 3 cm.

Freeze depth was measured on 11 days at CSSL and 5 days at the ski area (table 1). The layers thawed completely each day. The compacted snow froze an average of about 13 cm deeper than the uncompacted snow with a standard deviation of 6 cm. The frozen layer in the compacted snow tended to be about twice as thick as the frozen layer in the uncompacted snow.

DISCUSSION

Two simplified equations of heat flow and internal energy (Bengtsson, 1981; 1982) allowed estimation of the differences in thermal diffusivity and heat deficit between the compacted and uncompacted snow. When freeze depth after some elapsed time is known, thermal diffusivity may be obtained:

$$k_s = \frac{158.1\ S\ Z_f^2}{t\ p_s(-T_s)}$$

where k_s = thermal diffusivity (cm^2/s)
S = irreducible water content (cm^{3}water/cm^{3}snow)
t = time that net radiation ≤ 0 (s)
Z_f = freeze depth (cm)
p_s = snow density (g/cm^3)
T_s = snow surface temp. (°C)
158.1 = density of water (1 g/cm^3) x latent heat of melting (333.7 joules/g) x specific heat of ice (~2.11 joules/g °C at -1°C)

The irreducible water content was assumed to be 0.04, following Bengtsson (1981). Time was the number of hours (converted to seconds) that net radiation--measured by a Kahl allwave net radiometer mounted 1 m above an uncompacted snow surface--was less than or equal to zero. Snow density was estimated to the nearest 0.05 g/cm^3 from a combination of measurements with 500 cm^3 SIPRE tubes, Federal snow samplers, and isotopic profiling density gages. Snow surface temperature (T_s) was estimated from an empirical equation requiring net radiation (NR) and air temperature (T_a) at 1 m:

$$T_s(^{\circ}C) = -2.2 + 0.8T_a(^{\circ}C) + 3.3NR(\text{joules/cm}^2\ \text{min})$$

This equation was obtained by regressing snow surface temperature measured with an infrared thermometer against air temperature and net radiation for clear nights in April 1984 and April 1985 at CSSL. The equation is applicable only to CSSL conditions and should not be used elsewhere.

Thermal diffusivity was calculated for only 10 freeze depths at CSSL (table 1). Adequate input data were unavailable for one observation period at CSSL and for the ski area site. The calculated values for the natural snow ranged from 0.008 to 0.022 cm^2/s with a mean of 0.013 cm^2/s and a standard deviation of 0.004 cm^2/s. This mean value is at the high end of the range of reported values. While the equation is not particularly sensitive to reasonable changes in most of the variables, if the irreducible water content was changed from 0.04 to 0.02, the mean thermal diffusivity would drop to 0.006 cm^2/s. This value is in the mid-range of reported values. While some uncertainty persists in the inputs and consequently in the calculated values, inputs other than freeze depth and density were considered constant for both types of snow. Measurements of freeze depth and density were straightforward, and the small errors in these inputs should not affect the comparison between natural and compacted snow. For compacted snow, calculated thermal diffusivity ranged from 0.016 to 0.088 cm^2/s, with a mean of 0.051 cm^2/s and a standard deviation of 0.024 cm^2/s. This mean value is roughly four times that of uncompacted snow and much higher than any value in the literature. The high thermal diffusivity of compacted snow should be expected considering the accelerated densification and hardening due to compaction. Since conduction through ice grains is the primary means of heat transfer in snow, thermal conductivity (or diffusivity) varies with density and with the degree of bonding between adjacent grains (Mellor, 1964).

Cold content or heat deficit can also be estimated with a simple equation:

$$CC = Z_f p_s c_i \ (-T_s)$$

where CC = cold content ($joules/cm^2$)

c_i = specific heat of ice ($joules/g^\circ C$) (2.08 to 2.11 at -5 to $0^\circ C$);

other terms are as defined above.

This equation assumes a linear temperature distribution with depth and simply averages the surface temperature over the thickness of the frozen layer. Calculated cold contents ranged from 7 to 38 $joules/cm^2$ in the natural snow and from 24 to 114 $joules/cm^2$ in the compacted snow. The mean of the ratios of the cold contents of compacted snow to the corresponding cold contents of natural snow was 2.4. Equal surface temperatures between compacted and uncompacted snow were assumed, though a few temperature measurements and theory suggested that compacted snow should be 1 to $3^\circ C$ colder than natural snow. This temperature difference would increase the cold content difference.

The cold contents were converted to millimeters of melt equivalent by dividing by the latent heat of fusion (333.7 joules/g). Without accounting for the temperature difference, cold contents in the compacted snow were about 1 to 2 mm greater than in natural snow. A difference in cold content of this magnitude means that meltwater on the following day might not start to flow out of the freeze zone of the compacted snow for 1 to 3 hours after water was leaving the freeze zone of the uncompacted snow at the typical low rates of early morning snowmelt.

SUMMARY

Nighttime freeze depth was measured at two Sierra Nevada sites in both compacted and uncompacted snow. The depth of the frozen crust in the compacted snow averaged 13 cm greater than in the uncompacted snow or about twice as thick on a typical morning. Calculated thermal diffusivity at CSSL was about four times as great in the compacted snow as in the natural snow. Estimated cold content at CSSL was generally 1 to 2 mm of melt equivalent greater in compacted versus undisturbed snow. These characteristics of compacted snow in this limited set of observations suggest that snowmelt runoff from compacted snowpacks may be distributed in time differently than runoff from undisturbed snowpacks. Changes in runoff timing could be important when combined with other effects of snow disturbance and vegetation and soil alteration. The potential for such changes should be considered in assessing erosion and channel stability in areas subject to widespread snow compaction.

TABLE 1. Freeze Depth Measurements and Estimates of Thermal Diffusivity and Cold Content

Date	Freeze Depth(cm) U*	Freeze Depth(cm) C*	Density(g/cm^3) U	Density(g/cm^3) C	Thermal Diffusivity(cm^2/s) U	Thermal Diffusivity(cm^2/s) C	Cold Content(J/cm^3) U	Cold Content(J/cm^3) C
Central Sierra Snow Laboratory (CSSL)								
4-6	15	25	0.35	0.45	0.022	0.048	22	48
4-7	15	40	.35	.45	.014	.080	33	114
4-8	15	30	.35	.45	.014	.044	33	86
4-9	13	30	.35	.45	.013	.056	24	72
4-10	13	23	.35	.45	.018	.044	19	44
4-18	8	23	.40	.50	.011	.072	7	24
4-19	8	20	.40	.50	.008	.040	10	32
4-20	10	30	.40	.50	.012	.088	13	48
5-1	3	13						
5-5	13	20(to soil)	.50	.55	.009	.020	34	58
5-6	15	20(to soil)	.50	.55	.011	.016	38	70
Ski Area								
4-2	8	30						
4-9	8	20						
4-18	3	15						
4-24	0	5						
5-6	20	30						

*U = Uncompacted Snow C = Compacted Snow

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