EVAPORATION FROM A WINTER SNOW COVER
IN THE ROCKY MOUNTAIN FOREST ZONE

By

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The energy balance of the natural snow cover has been the subject of considerable study for the past two decades and has generated too large a literature to review in a short paper. Much of the work has concerned the wet or thawing snow cover which is of immediate interest to the practical hydrologist. The observations discussed in this paper concern a dry pack in the sense that no liquid or free water was evident by the simple tests available to the researcher and that snow temperatures remained below 0°C. with the exception of the first millimeter or so of the pack surface.

A fairly inclusive approach to the energy budget of the snow cover was made by Eckel and Thams (1).[2/] The total energy of the snowpack is equated to the heat content of the component ice. Factors neglected by this approximation include the sensible and latent heat content of the void air, the distribution of potential energy due to settling. On the assumption of a saturated void air, the first factor can be shown to be negligible for snow with densities in the range found in nature; the magnitude of potential energy changes can be shown to be quite small from measured settling and densities. On the basis of this approximation, the energy (H) of a unit column of the snow cover may be expressed as:

(1) $$H = \int_{o}^{z's} P_s \, C_i \, T_s \, dz'$$

The energy losses and gains by the pack over a given interval consist of absorbed solar radiation and long wave radiation loss from the snow surface, R; the gain of soil heat by the snow cover, S; the gain of sensible heat from the overlying air, F; and the associated gain or loss of latent heat,(p). In intervals with new snowfall, an additional term, W, must be considered representing the heat content of the new snow. The final energy balance is then:

$$\Delta H = R + S + P + F + W.$$

This balance allows an estimate of P to be made for a particular interval, as the residual of the remaining terms:

$$F = \Delta H - R - S - F - W,$$

or in terms of equivalent water lost, (E):

(2) $$E = \frac{W + F + S + R - \Delta H}{L}$$

where L is the heat of sublimation amounting to about 677 cal./gm.

The most difficult estimate involved in evaluating relation (2) for natural conditions is that of F. One commonly used method is to assume that the flux of sensible heat is due to the same small vertical motions responsible for transport of water vapor, in other words, that the eddy exchange coefficients (Km) for momentum and (kh) for sensible heat are equal. The validity of this assumption for a given situation has been shown

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to depend on the wind shear and temperature gradient as combined in the Richardson's number (Ri) (2):

where $Ri = \frac{g}{T} \left(\frac{\delta T}{\delta z} + \Gamma\right) \left(\frac{\delta \bar{u}}{\delta z}\right)^{-2}$ in the manner;

(2a) $Ri > -0.1;\ Km > Kh$

$Ri < -0.1;\ Km < Kh$

When $-0.15 < Ri < -.05$, the assumption holds to a fair approximation. Ri approaches this range for levels very close to the surface, even with large lapse rates, because of the rapid decrease in velocity at the surface. If the air layer considered is assumed to be in equilibrium with the surface, then the vertical momentum flux $Km \frac{\delta u}{\delta z} p$ will be equal to the drag force exerted on the flow by the snow surface (τ_o).

The low-level windspeed profile associated with a particular value of τ_o will be a function of the dynamic stability in terms of Ri and the surface Reynolds number Re* (3) where

(3) $$Re^* = \frac{u_* z_o}{\nu}$$

u_* above is the "friction velocity" defined as

(4) $$u^* = \sqrt{\frac{\tau_o}{p}}$$

and z_o is an effective surface irregularity dimension which will depend on the height of the actual physical protuberances and their average spacing over the surface. For

(5) $Re^* < 2.5$ ~~5.5~~; and $Ri = 0.1$,

the vertical variation of velocity with height follows the relation for smooth boundary turbulent flow;

(6) $$\frac{u}{u_*} = \frac{1}{k} \ln\left(\frac{u_* z}{\nu}\right) + 5.5.$$

Relation (6) is valid above the height (d) at which transport of momentum by molecular motions is negligible. For practical purposes, this height may be estimated as,

(7) $$d = \frac{11.6 \nu}{u_*} .$$

Given the assumption of stability and the consequent relations (2a),

(8) $$f = \tau_o \, C_p \left(\frac{\delta T}{\delta z}\right) \left(\frac{\delta u}{\delta z}\right)^{-1}, \text{ or}$$

(9) $$F = p u_*^2 \frac{\delta T}{\delta z} \left(\frac{\delta u}{\delta z}\right)^{-1} C_p .$$

A similar assumption may be made regarding the eddy exchange coefficient for water vapor (Kw) between levels above d. Observations reported in the literature (2, 4) indicate that this assumption is plausible over a wide range of stabilities. On this ground, the vapor flux may be estimated by

$$e = p\,(u_*)^2 \frac{\delta Q}{\delta z} \frac{\delta u}{\delta z}^{-1}$$

or by equation (9) above

(10) $$e = f \frac{\delta Q}{\delta z} \left(\frac{\delta T}{\delta z}\right)^{-1} \frac{1}{C_p}$$

Equation(10) may be used to estimate evaporation from the sensible heat flux if a local value of the gradient of the specific humidity (Q) and the associated temperature gradient are known. Relation (10) will be valid if $\left(\frac{\delta Q}{\delta z}\right)$ is measured above the level d;

otherwise, it will underestimate the vapor flux to the extent that the molecular diffusivity for water vapor exceeds that for sensible heat (4). At low temperatures and moderate windspeeds, if

$\frac{\delta Q}{\delta z}$ and $\frac{\delta T}{\delta z}$ are approximated by $\frac{Q2 - Q_o}{2}$ and $\frac{T_2 - T_o}{2}$, respectively,

$$e \simeq 1.1 \frac{Q_2 - Q_o}{T_2 - T_o} \frac{(f)}{C_p} \tag{11}$$

where Q_o is the saturation specific humidity over ice at the surface temperature T_o, and Q_2, T_2 are the specific humidity and air temperature at an elevation of 2 cm.

Under the assumptions made above, equation (2) may be used to evaluate E from the energy balance components including the sensible heat gain, F. E may also be estimated directly from the sensible heat flux and low-level temperature and humidity gradients by integrating equation (11) over the observation period.

The observations to be discussed were made at Fraser Experimental Forest at an altitude of about 9,000 feet in the Colorado Rockies. The site was a clearing about 500 feet in diameter at the foot of a wooded ridge to the south. From 0700 to about 1500 hours (LST), the site received direct sunshine. It was then shaded by a shadow cast by trees and a hill to the south of the site. Timber surrounding the clearing was primarily second-growth lodgepole pine to the east and north, which graded into mature lodgepole pine and a spruce-fir mixture to the west and south. In the second-growth timber, average height was 10-13 m. (35-45 feet) while in the mature stand heights ranged from 23-30 m. (75-100 feet). Height in the spruce-fir mixture varied extremely, from 1-30 m. (3-100 feet), typical of the uneven age distribution. The wind regime, typically light, was dominated by cold air drainage from the ridge during the night and late afternoon.

The time interval chosen for the observation periods was 12 hours, from 0600 LST to 1800 LST for daytime observations and 1800 LST to 0600 LST for nighttime observations; at these times, the snow cover temperatures were well below freezing at all levels.

The change in the snowpack heat content $\triangle H$ was calculated from the measured density profiles and the pack temperature profile as described in (1), at the beginning and end of the interval as indicated by equation (1).

The temperature profile was measured with a sequence of precision thermistors mounted on settlement disks (5, 6). One of these disks was dropped on the pack surface after each major snowfall during the winter, with a resultant spacing of about 10 cm. in a snow cover of about 80 cm. height by February.

Density profiles were taken with standard 500 ml. steel tubes inserted horizontally from a nearby trench. Surface layers were sampled by dropping a tube vertically through the upper 20 cm. of the snow cover. One such profile was taken during each sequence of observations.

The soil heat flux was estimated from the output of an Albrecht-type "heat flux" plate buried at 1 cm. below the soil surface. The voltage output from this plate was recorded automatically at 12-minute intervals. A set of thermocouples was buried adjacent to and beneath the flux plate; these were also automatically monitored to yield records of the soil temperature profile down to about 2 m. (6 feet). The ratio of the soil temperature gradient across the upper 20 cm. to the measured heat flux showed no variation within the accuracy of the measurements through the entire study period. It may be reasonably inferred that the thermal diffusivity in this region remained constant, and therefore no appreciable migration of moisture occurred across the soil-snow interface.

The total radiation incident on the snow surface and the difference between this flux and that reflected or emitted from the surface were measured by two Gier and Dunkle type thermal radiometers mounted at about 50 cm. above the snow surface.

Windspeed and shear were measured with a vertical array of sensitive anemometers on a mast at intervals of 20, 40, 80, and 160 cm. A scale was attached to the mast so that the elevation could be corrected for new snowfall. Situations for which near neutral stability in the sense of relations (3) prevailed above 50 cm. were infrequent for night-time periods. Such periods occurred normally during the late morning and during the afternoon, and fortunately with wind directions that indicated a uniform fetch of some hundreds of feet across snow surface, typical of the vicinity of the temperature array. On these occasions, the surface drag was estimated by extrapolation of the vertical wind-speed profile, which then showed a logarithmic variation. The apparent value for Z_o and the windspped indicated at 2.3 Z_o indicated surface Reynolds numbers less than 5.5 in all cases, and thus by criterion (5) the flow was smooth, conforming to relation (6). Since the low-level windspeeds were at a maximum for each observation on these occasions, and assuming no change in surface texture, it would follow that smooth flow prevailed through the period. The value of u_* for a particular value of Z and u may be thus computed by relations (6) and (9). The value of d calculated from the windspeed reached a maximum of about 1 cm. during early morning hours. u_* was calculated for each hour for the elevation of the lowest anemometer cup and the hourly average speed at that level. The measured difference in speed between the lowest cup, whose elevation varied from 10 to 20 cm. above the snow surface, and the second anemometer at an additional 20 cm. elevation, was used with the hourly average temperature gradients to estimate the flux of sensible heat to the pack surface.

Local air temperatures were measured at 2, 4, 8, and 10 cm. above the surface by a movable thermistor array and by a fixed Thermocouple at 125 cm. above the soil surface. Snow surface temperatures were measured by a thermistor at the bottom of this array. Temperatures at these points were automatically recorded at 12-minute intervals.

The energy contribution of new snowfall was computed from the accumulation on a snow board at the site, and surface temperatures during deposition. Accumulation height and density were measured at 4-hour intervals.

An attempt was made to measure low-level air humidity in order to compare evaporation losses computed by equation (11) above with those computed from the energy balance. A sealed chamber containing a diaphragm hygrometer was connected across the intake of a vacuum cleaner and an inlet tube leading from the heated instrument trailer to an intake fixed 2 cm. above the snow surface. At hourly intervals, the chamber was filled at a standard rate considerably less than the mass flow at the current windspeed, sealed, and allowed to reach equilibrium. The specific humidity of the air sample was estimated from the final hygrometer temperature and the indicated relative humidity. The arrangement functioned well at air temperatures above -10°C. Three satisfactory sequences were obtained which were used as the basis for the comparison mentioned below. $T_2 - T_o$ was evaluated from the temperatures recorded by the movable array.

The results of the energy budget calculated for the nine periods of observation are shown in Table 1.

Total evaporation was computed from relation (11) for 3 days. The computed values were 67 mg. cm.$^{-2}$, 46 mg. cm.$^{-2}$ and 200 mg. cm.$^{-2}$ for February 15 and 16 and March 22, respectively. The corresponding values from Table 1 are 59 mg. cm.$^{-2}$, 57 mg. cm.$^{-2}$ and 180 mg. cm.$^{-2}$. Instrumental errors for both series of measurements are estimated as plus or minus 10 percent due primarily to the limitations of the thermal net radiometers and the diaphragm hygrometer, respectively.

As is obvious from Table 1 and the bulk of the similar studies reported in the literature, the major energy component responsible for high evaporation losses is the net insolation gain, R. The estimation of the increase of total insolation with time through the season for a particular site is a fairly straight-forward problem if the local topography, the geometry of the vegetation, and the distribution of sky cover are known. More conjecture is involved in allowing for the gradual decrease in albedo. Lower albedos should be associated with an increase in average grain size and closer packing in the surface region of the snow cover (7). These changes may reflect both variation on the character of new snowfall and processes occurring in the deposited snow.

A few observations of the same type were made in the winter of 1962 (6). Similar calculations showed a general daytime evaporation of the same order of magnitude as the current data. There is however, a dramatic difference in the soil heat flux and the nocturnal evaporation rates. Soil heat contributions measured in the same manner as for the current data ranged as high as 40 cal. cm.$^{-2}$ in March as compared with the small values listed in Table 1. The temperature gradients through the snow cover and the soil diffusivity are of the same order of magnitude for both seasons. One feature present in the current snow cover was a 5 cm. ice layer caused by rain falling on the snow cover in January. This layer was about midway in the pack at the time of the observations, and was apparently of very low permeability. Such an obstruction to possible void-air convection was not present in the snow cover of the previous year. The relative impermeability of this layer was also suggested by the formation of depth hoar deposits on its upper surface, quite similar to those formed at the soil interface. If such a convection-blocking effect were in operation, it would be anticipated that for approximately equal thermal radiation loss at the surface, any nocturnal steady-state temperature distribution reached must involve a greater flux of sensible and latent heat to the pack surface and therefore higher condensation gains.

TABLE 1

Date	$\triangle$H	R	S	F	W	E	Albedo	Conditions
	cal cm^{-2}				mg cm^{-2}		Percent	
Feb. 13	- 7	- 37	+ 3	+13		- 21		clear night
Feb. 15	+ 6	+ 36	+ 3	+ 6	+1	+ 59	90	clear morning snowing in afternoon
Feb. 16	- 3	+ 31	+ 2	+ 3		+ 57	85	clear day
Mar. 7	-22	- 34	+ 1	+10		- 1.5		clear night
Mar. 9	+19	+ 94	+ 1	+12		+130	83	clear day
Mar. 10	-10	- 37	+ 2	+ 4		- 31		clear night
Mar. 12	+11	+ 60	+ 1	+ 3		+ 78	83	cloudy day
Mar. 22	+40	+146	0	+13		+180	73	clear day
Mar. 23	-28	- 48	- 1	+ 3		- 24		clear night

Comparison of the energy balance for three nights in March 1962 and the current March night period does show losses for the former of as much as 60 mg. cm.$^{-2}$ as compared with a gain of 22 mg. cm.$^{-2}$ for the latter. Consideration of such effects on the basis of the data presented here would be obviously quite speculative. It is hoped that a planned analysis of the hourly temperature profiles and the profile of air permeability through the snow cover will allow some definite conclusions.

NOTATION:

H = The heat content of a column of a base of 1 cm.^{2} extending to a height Z_s--cal. cm^{-2}.

Z'_s = The height of the snow surface above the soil-cm.

P_s = Local snow density--gm. cm.$^{-3}$.

C_i = Specific heat of ice (0.51)--cal. deg.$^{-1}$ gm.$^{-1}$.

T_S = Local snow temperature--deg. C.

Z	=	Height above snow surface-cm.
Z'	=	Height above soil surface--cm.
R	=	Net radiation gain if the snow surface per unit area is over 12 hours--cal. $cm.^{-2}$.
F	=	Net loss of sensible heat by the snow surface per unit area over 12 hours--cal. $cm.^{-2}$.
P	=	Net loss of latent heat by the snow surface per unit area over 12 hours--cal. $cm.^{-2}$.
S	=	Net gain of sensible heat at the snow-soil interface per unit area over 12 hours--cal. $cm.^{-2}$.
W	=	Net gain of heat content due to new snow per unit area of snow surface over 12 hours--cal. $cm.^{-2}$.
E	=	Water vapor lost by the snow cover per unit surface per 12 hours-- gm. $cm.^{-2}$.
Km	=	Eddy exchange coefficient for momentum--sec^{-1} $cm.^{2}$.
Kh	=	Eddy exchange coefficient for sensible heat--sec^{-1} $cm.^{2}$.
Kw	=	Eddy exchange coefficient for water vapor--sec^{-1} $cm.^{2}$.
Ri	=	Richardson's number.
u	=	Windspeed--cm. $sec.^{-1}$.
p	=	Air density--gm. $cm.^{-3}$.
τ_o	=	Surface drag force--dynes $cm.^{-2}$.
Re*	=	Surface Reynolds number.
Z_o	=	Surface roughness--cm.
ν	=	Kinematic viscosity of air--sec^{-1}.
u*	=	Friction velocity.
k	=	Von Karman's constant (0.4).
d	=	Height of laminar boundary layer above snow surface--cm.
C_p	=	Specific heat of air (0.240)--cal. $deg.^{-1}$ $gm.^{-1}$.
g	=	980 cm. $sec.^{-2}$.
Γ	=	Dry adiabatic lapse rate.
Q	=	Specific humidity.
f	=	Flux of sensible heat from snow surface--cal. $cm.^{-2}$ $sec.^{-1}$.
e	=	Flux of water vapor from the snow surface--gm. $cm.^{-2}$ $sec.^{-1}$.
Q_o	=	Saturation specific humidity over ice at T_o.
T_o	=	Snow surface temperature--deg. C.
ΔH	=	Change in snow cover heat content over 12 hours--cal. $cm.^{-2}$.

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