The SURFACE GEOMETRY of a closed conifer forest in relation to losses of intercepted snow

by Donald R. Satterlund
and Arthur R. Eschner

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RALPH W. MARQUIS, DIRECTOR

The Authors

DONALD R. SATTERLUND, formerly assistant professor of forest influences at the New York State University College of Forestry, is now assistant professor in the Department of Forestry and Range Management, Washington State University. He received his Bachelor's (1951) Master's (1955) and Ph.D. degrees (1960) in forestry from the University of Michigan. He has been teaching forest influences and watershed management since 1958. This paper was prepared at Syracuse, New York, while he was a member of the Cooperative Watershed-Management Research Unit of the Northeastern Forest Experiment Station and the New York State University College of Forestry.

ARTHUR R. ESCHNER, assistant professor of forest influences at the New York State University College of Forestry, received his Bachelor's degree from the New York State College of Forestry, received his Master's degree at Iowa State College in 1952, and joined the U. S. Forest Service in 1953. He has served in the Central States Forest Experiment Station, and in the Northeastern Forest Experiment Station at Kingston, Pennsylvania, and Parsons, West Virginia. This paper was prepared while he was stationed at Syracuse, New York, as the Forest Service project leader in the Cooperative Watershed-Management Research Unit of the Northeastern Station and the New York State University College of Forestry.

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INTRODUCTION

FROM snow studies too numerous to mention there is general agreement that much of the snow that accumulates on conifer forests is evaporated into the air and thus never contributes to water yields. Quantitative estimates vary widely, from 9 inches water equivalent for some seasons at the Upper Columbia and Williamette Basin Snow Laboratories (U. S. Army 1956) to negligible in other areas, depending upon vegetative density, quantity and frequency of snowfall, and many other factors.

Miller (1961), after a thought-provoking analysis of 110 investigations, suggests that snow research may be a victim of "technical folklore," and that snowfall interception may be grossly exaggerated. He questioned both the accuracy of snow measurements and the availability of energy required for snow evaporation.

The objective of the following analysis is to examine the process of snow interception in light of the involvement of geophysical and climatological processes within the peculiar geometry of snow held aloft by a conifer canopy, in contrast to the same processes acting on a snowfield on open, level ground. Perhaps an analytical contribution of this kind will stimulate research that will lead to a better understanding of snow interception, and ultimately to application in forest-land management for water yield.

ENERGY SOURCES FOR EVAPORATING SNOW

Because snow-interception losses involve evaporation to the atmosphere, and evaporation requires energy (more than 600 calories per gram of snow converted to water vapor) we must first consider the sources of this energy. They are essentially three: (1) the originating source, solar radiation; and the secondary sources, (2) long-wave radiation and (3) advected sensible heat.

Solar Radiation

Solar radiation may supply considerable quantities of heat. For instance, prior to the melt season of 1946 in the Sierra Nevada up to an average of 350 langleys was received daily, and during the melt season in April and May, 650 to 700 langleys (Miller 1950). The latter values were associated with a net radiation balance of about 100-200 langleys daily, which would cause evaporation of 0.2 to 0.4 grams of snow per square centimeter if all were used in evaporation. However, much is often used for melting snow. Yet even 0.15 grams of snow per square centimeter evaporating daily amounts to 1.77 inches per month, certainly not a small amount.

The impact of solar radiation on intercepted snow in a closed conifer stand may be quite different than on an open snow-covered area. The irregular topography of the intercepted snowfall results in concentration, dispersion, and shading of the solar radiation. The approximate pattern of direct-beam radiation can be deduced from a simplified model of a conifer crown as a cone, and a stand as a series of cones overlapping at the base. Taking any given conditions of spacing, mean height of upper crown closure, and height of crown tip above level of closure, and height, diameter, and thus slope of the cone surface can be defined. Dividing the cone surface into eight aspects permits calculation of the approximate quantity or proportion of the direct-beam radiation received at any given time on each slope and aspect, by means of procedures outlined in Fons *et al.* (1960). In a closed stand, the area of shadow would also require determination.

In this geometrical situation the implications of the concentration of radiation to snow evaporation, especially under conditions of prevailing air temperatures below freezing, are clear. Where radiation is intense on the snow surfaces, strong vapor pressure gradients could develop from snow to air, causing considerable evaporation; whereas, if radiation were dispersed over a level snowpack, the same quantity of radiation per acre might be insufficient to cause evaporation. Qualitative evidence of this phenomenon has been observed by the authors on days when the maximum air temperature did not exceed 20°F., in late December when direct-beam radiation was at its weakest.

Albedo

However, the intensity of incoming radiation at any given point on a snowfield is not a direct measure of the energy absorbed. A portion of the radiation is always reflected; the proportion is the albedo of the snowfield. The amount absorbed is the complement of the albedo, and absorption is the significant element of the snowpack radiation balance, for only energy absorbed can contribute to melt, evaporation, or heating of snow. For example, an albedo change from 90 percent to 80 percent is only an 11-percent reduction in reflection, but it represents a 100-percent increase in absorption. This distinction should be kept clearly in mind when assessing the significance of albedo, particularly where high albedos are involved.

The albedo of an intercepted snowfield, in contrast to that of a snowfield on open ground, is strongly influenced by the opportunities for multiple reflection. For instance, where snow is essentially continuous over a conifer canopy, escape of reflected radiation would be at its maximum, but reduced to the extent that some is reflected to snow on the sides of adjacent trees. When snow is discontinuous, a far more common situation, considerable reflection to the underside of snow-covered branches and into the snowless portions of the trees occurs. Thus an albedo of 90 percent, measured in the open, may be reduced in the intercepted snowfield by multiple reflection to 81 or 73 percent, or less, and almost completely if reflected onto snow-free surfaces. Absorption is, commensurately, 10, 19, or 27 percent or more in the canopy snowfield.

The fact that the depth of the intercepted snowfield is generally much shallower than depth in the open will also reduce the albedo. Kuzmin (1956) indicated that the albedo of a continuous cover of snow at the beginning of a period of melt (not throughout the period) decreased from 82 to 60 percent as the depth of the snow varied from 80 to 5 cm.; the latter figure a fairly representative depth of intercepted snow, the former a reasonable snowpack depth in many areas. The amount of radiation absorbed by the thinner snow is more than double that of the thicker layer, even though albedos differ by only 22 percent.

The major reason for initial differences in albedo of snow similar in all respects except depth is that snow reflects radiation not only from the surface, but also from underlying layers (Kuzmin 1956). It is well known that snow transmits considerable quantities of radiation: as much as 50 percent of the radiation that penetrates the surface may reach a depth of 10 cm., and 10 percent may reach a depth of 30 cm. (Geiger 1959). Molchanov (1960) indicated that a snow layer 3 cm. thick, and having a density of 0.16, transmitted about 32 percent of the incident solar radiation, but a layer of 35 cm. of snow let through only 1.5 percent. In a deep snowpack, only part of this attenuation would be due to absorption; the other part would be due to internal reflection that escapes from the snowpack. If we consider an intercepted "snowfield," shallow in depth and overlying dark coniferous foliage of low albedo, it becomes clear why the thin "snowfield" has a lower albedo than the thicker one.

The effect of thickness of the intercepted "snowfield" on the quantity of incoming radiation absorbed is not limited to initial differences in radiation absorption as developed above. An intercepted snowfield 5 cm. deep and having a density of 20 percent has a mass of 1 $g./cm.^2$ With a specific heat of 0.5 cal./g., net absorption of only 0.5 cal. would provide enough energy to raise the temperature of this thin snowpack by 1°C. A deeper snowpack on the ground absorbs radiant energy through a greater depth, and therefore a smaller portion is absorbed in the upper 5 cm. layer. so it heats more slowly. Since a very thin layer of water normally covers ice or exists in snow crystals at temperatures somewhat below freezing (Chalmers 1959), the surface of the intercepted "snowfield" becomes wet sooner. The effect of increasing the liquid water content is to reduce albedo, and thus an even greater differential in capturing incident radiation begins, until the final disappearance of the intercepted "snowfield".

Sensible Heat

Sensible heat of the atmosphere appears as a possible important source of energy for snow borne on the canopy of a conifer forest. If the air is warmer than the snow, some of this energy can be transferred by turbulent exchange to the snow. Even if no temperature gradient exists initially, and snow and unsaturated air freely co-exist at the same temperature, nothing can prevent vapor transfer from snow to air as long as a vapor pressure deficit exists in the atmosphere. Energy for the vapor transfer is derived from the entire system; the sublimating snow becomes cooler, a temperature gradient develops between air and snow; and the whole system cools until both temperature and vapor pressure are in equilibrium.

Sverdrup's equations for turbulent heat and vapor transfer to or from a snow surface, as presented by Diamond (1953) are:

$$O = \frac{Cp\ Pk_o^2\ Va\ (t - t_s)}{\ln\ (a/z_o)\ \ln\ (b/z_o)} \qquad [1]$$

$$F = \frac{0.623}{p}\ \frac{Pk_o^2\ Va\ (e - e_s)}{\ln\ (a/z_o)\ \ln\ (b/z_o)} \qquad [2]$$

where O = heat transfer in $cal./cm.^2/sec.$; F = evaporation in $g./cm.^2/sec.$ (cm./sec.); Cp = specific heat of air at constant pressure, 0.24 cal./g./°C.; P = density of air, 1.276×10^{-3} $g./cm.^3$

at 1,000 mb. and 0° C.; V_a = wind velocity at height a, cm./sec.; t = air temperature at height b, °C.; t_s = snow surface temperature, °C.; e = vapor pressure of air at t, mb.; e_s = vapor pressure of snow at t_s, mb.; k_o = von Karman's constant, 0.42; z_o = roughness parameter, 0.25 cm. for snow from Sverdrup (1936); p = surface pressure, assumed 1,000 mb.; b = height of temperature and humidity element, assumed 2 m. to approximate height of instruments in shelter; a = height of anemometer, assumed 6 m.

It is assumed in this study by the use of the Sverdrup model that eddy diffusivity of heat and vapor are the same as for momentum transfer. Portman *et al.* (1961) indicate that this relation is reasonable when lapse conditions do not depart too far from the adiabatic, but it is questionable for other lapse conditions. Brooks (1959) suggests that eddy diffusivity may be considerably less under inversion conditions and considerably more under thermal convection.

With these limitations in mind, a review of equations [1] and [2] reveals possibilities for increased turbulent heat transfer to intercepted snow in a forest canopy as contrasted to snow in a level field. It is possible that differences in the measured variables of wind speed, temperature of air and snow, and vapor pressure may exist between a "snowfield" in a canopy and one on open ground.

If the reference elevation from which a and b in equations [1] and [2] are measured is the snow surface in the case of the open, level snowfield, then the reference elevation of the "snowfield" held in the canopy would approximate the elevation of complete crown closure (Baumgartner 1956). How would the above variables be affected when measured above the "snowfield" of the crowns as contrasted to an open, level snowfield?

Wind speed measured at 6 m. above our reference elevation is likely to show some change. Baumgartner (1956) indicates that above a forest of limited extent to windward wind was stronger than in the open, because the higher strong winds from the open were impressed on the layers above the tree tops. When wind was measured after passing over an extended area of forest from another direction, the differences were less. Wind speed, therefore, may be slightly greater over the forest.

Comparative data for temperature and vapor pressure at 2 m. above the reference elevation of a crown "snowfield" and that in the open were not found for winter conditions. It will be assumed that differences are negligible. During all periods when the air temperature equals or exceeds 0° C., it can be assumed that the temperature and vapor pressure of the snow surface in the open field or in the canopy should be the same, as snow cannot assume a temperature in excess of 0° C., and unless a strong inversion is present, it will quickly assume its maximum temperature. When air temperatures are less than 0° C., snow surface temperatures in the crown canopy are likely to slightly exceed those on the ground in the open, for reasons to be discussed later.

With regard to the remaining elements in equation [1]; specific heat of air and von Karman's constant should be the same. Density of air would be only slightly decreased, depending on the height of the stand, and will be assumed negligible. Only the roughness parameter is likely to be substantially different when an intercepted "snowfield" is compared with a snowfield on open, level ground.

Of all the elements in equations [1] and [2] the possible differences in roughness parameter exhibits the most fruitful possibility for explaining increased evaporation from an intercepted "snowfield" in a closed conifer forest as contrasted with a snowfield on level, open ground. It is obtained by plotting wind speed at several elevations above a given surface against height; the curve is extrapolated to zero wind velocity, and the height intercept denotes the roughness parameter of the surface (Light 1941).

Wind velocity in relation to height above a surface follows a power law (Baumgartner 1956, Brooks 1959). By plotting wind speed above a forest canopy during periods of neutral stability, we can determine the roughness parameter for a given closed conifer stand if we can determine a suitable reference surface. If we accept Geiger's (1959) concept of an outer active surface for a forest canopy, we cannot use the ground surface as our zero plane, or the height of the surface above which roughness elements project, for then it would automatically follow that the roughness would be a function of total tree height. Brooks (1959) presents a means whereby the zero plane displacement can be calculated, and above which the roughness parameter would be measured, as does Baumgartner (1956). The zero plane displacement is upward into the approximate level of full crown closure. Using data from a young dense pine plantation in Germany, Baumgartner (1956) calculated a z_0 of 290 cm. Byers (1959, p. 515) presents a formula from which he calculates the roughness parameter:

$$\frac{U_1 = \ln (z_1 + z_0)/z_0}{U_2 = \ln (z_2 + z_0)/z_0} \qquad [3]$$

where U_1 and U_2 = wind speeds at height Z_1 and Z_2, respectively; Z_1 and Z_2 = anemometer height above a given surface; and Z_0 = roughness parameter. He states that with a park-like distribution of trees, Z_0 is about 100 cm., but the source of his data is not given.

Rough calculations by the authors from wind speeds above and in the crown space of two 65-year-old Scots pine stands (data from Geiger 1959) and a 50-year-old dense ponderosa pine stand (data from Fons 1940) by the graphical method and equation [3], using assumed zero plane displacements, yielded a range in roughness parameters from 30 to 175 cm., with a mean of about 80 cm. However, these figures are probably conservative, as the anemometer placement did not extend more than 3 m. above the stands in any case, and a true free-air profile was not ascertained. Estimates of roughness parameters therefore range from 30 to 290 cm., a surprisingly broad range, considering that the latter is associated with the most uniform crown canopy. For a level open snowfield, Sverdrup (1936) gave a roughness parameter of 0.25 cm. and Geiger (1959) a figure of 0.5.

From equation [1] we can estimate the effect of changing roughness parameters by letting the numerator = K, a constant for any given air mass and snow surface temperature.

Completing the equation with z_0 = 0.50 cm. for a smooth snowfield, we get heat transfer to or from it in the quantity K/43.03816. If we choose a roughness parameter for the forest of 100 cm., heat transfer to the intercepted snowfield equals K/1.24195, or nearly 35 times that to the snowfield on open level ground. If we choose the lowest roughness parameter, 30 cm., our equation reduces to K/5.68475, or about 7.5 times the heat transfer to an open level snowfield. Any forest parameter greater than 200 cm. gives a nonsense answer of heat transfer against the temperature gradient. Either the roughness parameter of 2 m. is too great, or equation [1] is invalid for such high roughness, or both.

Thus the effect of surface roughness can be substantial, if the other elements of the equation remain constant. However, it is difficult to conceive that initial temperature gradients between air and snow could long continue to exist under such conditions of high wind and roughness parameter. Wilson (1941), assuming a value of 700 m. for the thickness of the turbulent layer, 0.0012 for air density, and a heat loss of 10 cal./cm.^{2}/hr., has calculated the loss of heat in an air mass moving 10 m. per second as about 0.5° C. in 25 miles over snow. Unless a snow-covered area were near a heat source, such as the sea, or on a lee mountain slope subject to adiabatic warming, the temperature of the air mass would soon drop to a level in which no amount of turbulent action could transfer appreciable sensible heat (Miller 1963, personal communication).

Even if the constant, K, were positive (indicating heat transfer to snow) rather than negative (indicating heat transfer from snow), as the net of accumulated advective heat transfer for the snow season should be in mid-latitude locations, it is obvious that turbulent heat transfer from air to snow would account for large interception losses only if the energy were used for evaporation rather than melt.

The vapor transfer equation [2] is analogous to equation [1] for heat transfer, but whether condensation (with or without melt) or evaporation occurs is dependent upon how the vapor pressure gradient is synchronized with the temperature gradient. In moist maritime air, evaporation is not likely; but evaporation situations can be found in the Chinook of the West and weakly in the subsiding air following the passage of a cold front.

Muller (1961) has calculated net snow evaporation for an open field at Syracuse, N. Y., for February 1961. Mean daily temperature ranged from −7° F. to 47° F. Evaporation occurred for 521 hours during the month, and condensation 151 hours. Net evaporation occurred on 23 of the 28 days and amounted to 0.296 inches of water equivalent. During this period, there were only 294 hours of possible solar radiation. Further, condensation occurred for only a maximum of 100 hours during 378 hours of nocturnal conditions. If this month is typical, then atmospheric conditions suitable for turbulent heat transfer to snow synchronized with vapor transfer from snow must be present nearly one-half the time, neglecting entirely the possibility of a warmer snow surface temperature from solar radiation as a cause of evaporation. It should be pointed out, however, that a weak foehn occurs at Syracuse (elevation about 400 feet) whenever the wind direction has southerly component, from whence elevations approximate 1,200 to 2,000 feet within 20 miles; but these conditions are common throughout much of the Northeast. These data suggest that, had snow been in a conifer canopy during this entire period, interception losses could have reached 2 or more inches water equivalent.

Persistence of Snow in a Conifer Canopy

Obviously snow does not persist continuously in a conifer canopy. Observations by Lull and Rushmore (1961) in the northern Adirondacks indicated that snow was present in various amounts on branches, nailed at a height of 5 feet on a wooden post, two-thirds of the time from December 8, 1959, to March 22, 1960. Wind velocities in the upper canopy of a stand might easily be higher than at the point of observation, reducing this — by blow-off and increased rates of evaporation and/or melt — by an unknown amount.

Twenty-two systematic observations by the authors from November 30, 1961, to April 27, 1962, at Tully Forest, New York, revealed intercepted snow in the tree crowns 63.6 percent of the time. The entire winter was deficient in quantity and frequency of snowfall, with only one snowfall in the entire month of March.

It seems reasonable that snow in central and northern New York may be available for interception loss for one-half the period of observation, during which snow on the ground is continuously present. Kittredge (1953) indicates similar relative persistence of intercepted snow in the Sierra Nevada Mountains.

Considering that the increased magnitude of energy supplied by turbulent transfer of heat from air to snow considerably exceeds the reduction in availability of snow for evaporation from a closed conifer stand as contrasted with a snowpack on level, open ground, we may conclude that the surface geometry of the conifer stand, expressing its effect on the roughness parameter, is sufficient to account for considerably higher losses from an intercepted "snowfield".

The above assumes that wind speed above and snow surface temperatures on the intercepted "snowfield" are the same as those on open, level ground. Both, however, are probably somewhat higher in the canopy, which would increase intercepted losses in relation to losses from an open, level snowpack. The explanation for increased wind speeds was given previously, but the increased temperature requires explanation.

A forest canopy, located at a greater or lesser distance above the ground, is always more or less porous, even where most of it is covered with snow. Theoretically, this should affect the formation of any inversion layer to a different degree than would occur above a snowpack on level, open ground. Air that is cooled in contact with cold snow becomes more dense. On the ground, it can only stagnate and build up a strong, if shallow, inversion. In the canopy, however, this cool air can sink through the porous layer, being replaced with slightly warmer air. Or, it can flow off the canopy into surrounding open areas. Geiger (1959) mentions a distinct forest breeze, detectable at a considerable distance from the forest edge, during a calm night with strong outgoing radiation. The subsiding air could possibly add small quantities of heat to the snow on the canopy, thus maintaining a slightly higher temperature, while lower temperatures are created on the snow in the open. A practical application of this phenomenon is being tested to increase the persistence of the snowpack in clearcut areas in the forest snow zone of California (Anderson 1963).

Long-Wave Radiation in the Canopy

Not all radiation absorbed is available for evaporation of melt in snow. All bodies continuously radiate heat, according to the Stefan-Boltzmann law, in an amount proportional to the 4th power of their absolute temperature. Little difference should exist in absorption of atmospheric or cloud long-wave radiation between the two snowfields considered, for snow is essentially a black body to most long-wave radiation and the difference in albedo would be negligible. Similarly, during periods when the air temperature is 0° C. or above, both should radiate at the same rate per unit area, as snow cannot attain a temperature higher than 0° C.

When air temperatures are less than 0° C., on clear nights and under conditions of incoming solar radiation, intercepted snow should radiate at a slightly greater rate than snow in an open, level field, because of the higher temperature of the snow surface indicated previously. The mean increase in the temperature of the intercepted snow is unknown, so the increase in outgoing radiation can only be surmised.

Snow as an Energy Storage Factor

The capacity of snow to store energy as sensible heat has been referred to in the discussion of the effect of thickness on the albedo of the contrasting snowfields. Another factor that influences the disposition of absorbed energy is the capacity of the snow to retain liquid water, and the disposition of melt water not retained in the snowpack.

Melting and evaporation of snow can occur over a surprisingly broad range of temperatures (Diamond 1953, Chalmers 1959). If liquid water is retained in the snow, only to refreeze later, melting represents a heat sink with a capacity of 80 cal./g. (the latent heat of fusion of water), while freezing represents a heat source of the same capacity. A thin snowpack held in the crown canopy of our closed conifer forest would be unlikely to retain more than 10 percent of its original water equivalent, considering that the melt process in itself reduces the depth of the snow. For a snowpack of 1 g./cm.^{2}, this would amount to only 0.1 g., the melting of which, and later refreezing, would convert only 8 calories of heat. On the other hand, a deep snowpack could not only retain a considerably greater quantity of water, but by gravity drainage could distribute it throughout the snowpack.

This differential in the two contrasted snowfields is important, for it is possible to carry over energy received during periods of incoming energy that is available as a source of energy for evaporation or outgoing long-wave radiation at night. Energy so carried over and lost to outgoing radiation is obviously not available for evaporation and must be deducted from incoming energy to obtain the amount available. Miller (1950) has estimated that the carry-over may represent 100 langleys daily in the snowpack on the ground in the Sierra Nevada mountains in early May.

A major implication of the limited capacity of intercepted snow as a heat sink may be of considerable importance in causing evaporative losses. If the carry-over energy capacity is only a small quantity, such as 8 or 10 cal./cm.^{2}, then only very short periods of a positive radiation balance will result in snow evaporation and/or melt — especially considering that intercepted snow is often held in positions more normal to the sun's rays than snow on open ground. The larger heat sink represented by deeper snow on the open ground would require a much longer period of radiation surplus before similar evaporative conditions are developed.

For example, assume an albedo of 70 percent for each snowfield; a heat-sink capacity of 10 cal./cm.^{2} for snow in the crown and 50 calories for snow in the open; and consider that the snow in the crown is oriented from 0° to S 45°, while that in the open is horizontal (0°). Calculations of the approximate direct-beam radiation absorbed in the two snowfields between 11:00 a.m. and

1:00 p.m. on an average clear day in December at latitude 42° N. range between 11 (0° slope) and 26 (S 45° slope) cal./cm.² This would provide a surplus of up to 16 calories available to evaporate exposed snow in much of the crown and none in the snowpack in the open. The daily radiation balance would be negative.

This small difference, accumulated over the snow season, could prove substantial. Therefore the capacity of snow as a heat sink may be critical in determining the length of the period of radiation necessary to cause snow evaporation. A daily radiation balance may be sufficiently accurate to account for losses from a deep snowpack on open ground, but frequently misleading if used to account for losses from intercepted snow.

Consideration of melt water not retained in the snowfield further increases the disparity in evaporative losses within a forest canopy and on snow-covered open ground. Except during the final stages of disappearance of the snowpack on the ground in spring, most melt water drains into the soil or at least below the snowpack. It becomes essentially unavailable for evaporation, for to get evaporation, energy must be applied to the water, which is protected by the snowpack above. If snow in the crown begins to melt at a rate exceeding the "snowfield" storage capacity, drip begins. Miller (1955) mentions that drip is exceedingly common in the Sierra Nevada, but observations by the authors in central New York indicate that while drip is not exceptionally rare, neither is it common. This drip often wets much of the crowns of trees, and maintains them in a wet condition for long periods. The wet surface reduces the albedo of the crowns; liquid water films may warm considerably — thus breaking the 6.11 mb. vapor pressure barrier of snow — and evaporate under conditions that may result in condensation on snow in the open.

During advection melts, air temperatures are often relatively high. Heat is available and considerable evaporation from tree crowns must occur in many instances, if only from the intercepted drip. If radiation melt occurs during clear weather and freezing temperatures, evaporation from melt water in tree crowns might even be greater. The combined evaporation from snow and retained melt water in the forest canopy may exceed evaporative losses from a snowpack on level ground even more than during non-freezing periods. Goodell (1959) measured the loss of snow from a severed branch suspended on scales in full sunlight at Fraser, Colorado. Air temperature ranged from −7° to +5° C., the dewpoint was $-14°$ C., and there was little wind. In just 2 hours and 40 minutes, 128 grams of snow were lost by evaporation and sublimation. Melt, but no drip, occurred. These data are sufficient to indicate that intercepted snow can be lost rapidly.

Canopy Coverage as a Factor in Evaporation of Intercepted Snow

The pattern of incident radiation, differences in albedo, thickness, capacity for heat storage, and disposition of melt water, all tend to accelerate the rate of exposure of bare crown canopies following snowfall, whereas snowfields on the ground remain essentially continuous throughout the snow season.

As more and more bare foliage appears, more solar radiation is absorbed, with consequent increased foliage temperature above ambient air temperature. Ehlers (1915) indicated that needle temperatures may exceed air temperatures by as much as 9° C. on a clear winter day in Ann Arbor, Mich. (latitude 42.5° N.). Sensible heat transfer by advection and convection, and long wave radiation, could provide appreciable additional energy to snow patches that remain. As the amount of snow cover decreases, the rate of heat transfer, and thus evaporation and/or melt of intercepted snow, should increase until it finally disappears. This process, frequently repeated in the forest canopy, would be largely inoperative during most of the snowpack season in an open, level field.

DISCUSSION AND CONCLUSIONS

All factors considered, there seem to be great differences in the energy and vapor balance between an intercepted "snowfield" and one on level, open ground which appear to result from the complex surface geometry of a closed conifer forest. There is little question that these differences are sufficient to account for considerably greater losses from intercepted snow. It would be tempting to present a quantitative comparison of the water balance between the two contrasting situations, but too few data are available, and interactions are too complex to provide other than speculation.

Despite the limitations in the assumptions, lack of data, and ignorance of energy transfer in complex situations, this analysis has revealed areas that should receive greater cognizance in snow-interception studies. Many problems of advection and turbulent heat transfer to snow in a forest canopy have received scant attention. The potential water losses attributable to this energy source are sufficient to deserve more intensive investigation.

Radiative heat fluxes are better understood and have received wide attention, but under the complex situation of snow in a forest canopy much remains to be done. Questions of depth, location, and persistence of intercepted snow, canopy exposure, structure, and storage capacity are little known but clearly pertinent to our quest. And rigorous means of characterizing forests in relation to these factors are largely lacking.

The classical type of snow-interception study, whereby snow is measured on the ground under different types and densities of vegetation, has yielded much useful information. For continued yield of new knowledge, however, such studies have probably passed the point of diminishing returns under most forest conditions. Snow losses should be studied where they occur — in the trees themselves. Combined with the classical interception study to serve as a quantitative check, a better understanding of the process of snow interception should result.

This qualitative analysis of the loss process suggests that snow interception losses may not be so grossly exaggerated as they appear to be.

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