SNOWMELT IN HARDWOOD FORESTS

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ABSTRACT

Snowmelt differences among forest covers result from differences in energy transfers by solar and longwave radiation, convection, and condensation-evaporation. Absorption of solar radiation is variable, but it can be estimated by using a canopy model. During the day, downward longwave radiation is about 0.04 cal $cm^{-2}min^{-1}$ higher under hardwood forest than in the open. Air temperature and humidity have practically the same values under a hardwood canopy as they do in an adjacent cleared area, while wind speed is reduced under hardwoods to about half that in the open. Any differences in convection-condensation melt between hardwoods and the open would arise from the difference in wind speed and from possible differences in turbulent transfer parameters. Transport of melt water through the soil to a stream can be modelled by an electrical analog. Snowmelt rate thus may be estimated from measured streamflow.

SNOWMELT in hardwood forests differs from melt in the open and in coniferous stands. The snow surface is partly but not completely shaded. Longwave radiation comes to the snow both from the sky and from the canopy. Air temperature, humidity, and wind near the snow surface may differ from these atmospheric properties above the canopy.

Hardwood forests are intermediate between conifers and open areas with respect to snowmelt. Fifty-five percent of New England is covered with hardwood forests, yet there has been little study of how these forests affect snowmelt.

In the Northeastern Forest Experiment Station we are trying to fill this gap in order to be able to predict how any change in forest land management will affect snowmelt and floods. In this paper we will tell you what we know now, what we do not know, and what we are trying to find out.

Snow generally melts fastest in the open and slowest under conifers with hardwoods in between (Maule 1934; Lull and Rushmore 1960; Hart 1963). These differences can be attributed to differences in shading, longwave radiation, wind speed, air temperature, and humidity. All these factors influence the gain and loss of energy by the snow cover. Since snowmelt requires energy for the phase change of water, energy-budget concepts are useful for studying snowmelt. However, the relative energy budgets of the three cover types have not been quantitatively and theoretically analyzed. Until we know how energy budgets are affected by vegetative cover, we will not be able to quantitatively predict melt-rate changes and thus changes in flood potential due to alteration of the forest cover.

Energy Balance of Snowmelt

There are three major sources of heat for snowmelt: radiation, convection, and condensation; and two minor sources, soil heat and heat stored in rain. In the Northeast, when rapid melt is occurring, solar radiation, convection, and condensation all contribute importantly to melt in the open, while net longwave radiation loss is very small (Federer 1968). Groundmelt is important in maintaining midwinter streamflow (Federer 1965), but it is negligible in the snowmelt season. Rain melt probably becomes more significant farther south, but we have found it to be very small at the Hubbard Brook Experimental Forest in central New Hampshire (Federer 1968). We need to analyze Weather Bureau records for stations scattered over the Northeast in order to determine the significance of rain melt once and for all.

Ignoring ground melt and rain melt, we write an equation for the energy budget of a snowpack as

$$\underbrace{L_f M}_{\text{melt}} = \underbrace{\left[(1 - a) R_{sd} + R_{ld} - \sigma T_o^4\right]}_{\text{net radiation}} + \underbrace{\left[C_p \rho K u (T_a - T_o)\right]}_{\text{convection}} + \underbrace{\left[L_v \rho K u (q_a - q_o)\right]}_{\text{condensation}} + \underbrace{S}_{\text{heat storage}}$$

in which:

L_f = heat of fusion of water (80 cal g^{-1})

L_v = heat of vaporization of water (595 cal g^{-1})

σ = Stefan-Boltzmann constant (8.11×10^{-11} cal $cm^{-2} min^{-1}$ $^{o}K^{-4}$)

ρ = density of air (1.29×10^{-3} g cm^{-3})

C_p = specific heat of air at constant pressure (0.25 cal g^{-1} $^{o}K^{-1}$)

a = albedo of snow surface (dimensionless)

M = melt rate (g cm^{-2} min^{-1})

R_{sd} = downward flux density of solar radiation at snow surface (cal cm^{-2} min^{-1})

R_{ld} = downward flux density of longwave radiation at snow surface (cal cm^{-2} min^{-1})

T_o = surface temperature of snow (^{o}K)

T_a = temperature of air some height above snow (^{o}K)

q_a = specific humidity of air at same height (dimensionless)

q_o = specific humidity at snow surface (dimensionless)

u = wind speed at some height above snow (cm min^{-1})

K = turbulence parameter (dimensionless)

S = change of heat storage in snowpack (cal $cm^{-2}min^{-1}$)

Net radiation includes four fluxes: downward and reflected solar radiation and downward and emitted longwave radiation. The fraction of downward solar radiation reflected by the surface is the albedo, a. The net solar radiation absorbed by the surface is thus (1 - a) R_{sd}. Since the absorptivity and emissivity of a snow surface for longwave radiation is nearly unity, all downward longwave radiation is absorbed, and the upward longwave flux depends solely on surface temperature, σT_o^{4}.

The signs of the convection and condensation terms are positive when heat is supplied to the surface, but these terms may change sign if the air is cooler or drier than the surface. The condensation term then becomes evaporation. For melting snow, the surface temperature is fixed at 0°C, and the surface specific humidity is fixed at 3.8×10^{-3}.

Separation of the heat storage term, S, and the melt term, $L_f M$, is complicated. We let S represent changes in the cold content of the snowpack. When S is negative, the pack is cooling either by freezing of liquid water held internally or by decreasing its temperature, or both. When S is positive, the pack is either warming or melting ice to the point when the pack becomes both isothermal at 0°C and wetted sufficiently that further melt causes water to drain from the pack. True melt, M, occurs only after this point and produces water that can leave the pack. Either S or M must be zero at all times. Rapid cooling or slow percolation may allow drainable water to refreeze before it can leave the pack; this freezing and subsequent thawing is included in M. However for the shallow low-density packs of the Northeast the amount of water affected in this way is negligible (Gerdel 1953).

Solar Radiation in Leafless Hardwood Forests

Incident solar radiation above the forest canopy is the same as in nearby open areas. But when the sun is shining, the stems, branches, and twigs of the leafless hardwood canopy produce an intricate network of light and shadow on the snow surface. The disposition of solar energy by canopy absorption, reflection, and snow absorption is crucial for the energy balance of snowmelt. This disposition varies with slope and aspect, solar azimuth and zenith angle, sky characteristics such as cloudiness, canopy characteristics, and ground or snow surface reflectivity.

To make some sense out of all these variables we undertook to develop a theoretical mathematical model for solar radiation in a leafless hardwood forest (Federer 1971). In the model the canopy is divided into two parts, a stem space of randomly arranged vertical cylinders and a crown space that is a uniformly absorbing medium. At present the model works only for horizontal surfaces; slope, aspect, and solar azimuth are not included.

The transmission of the canopy for solar beam radiation turns out to be

$$t = \exp(-P_c \sec\theta - P_s \tan\theta)$$

where θ is the solar zenith angle, P_c is a crown space parameter that must be found empirically to fit field data, and P_s is a function of stand parameters.

$$P_s = 2\,Bh / \pi D$$

where B is the fractional basal area, h is tree height, and D is the quadratic mean diameter of the stems.

The fraction of the downward solar radiation that is beam radiation is called F, and so the diffuse or sky fraction is 1 - F. Canopy transmissivity for diffuse radiation, t_d, is t integrated over a uniformly radiating hemisphere. Total canopy transmissivity is

$$T = t\,F + t_d\,(1 - F)$$

Then

$$\text{Snow absorptivity} = (1 - r)\,T$$

where r is the reflectivity of the snow. And

$$\text{canopy absorptivity} = (1 - a)\,(1 - T\,(1 - r\,(1 - t_d)))$$

where a allows for reflection from canopy surfaces. A value of 0.10 for a seems empirically appropriate. The albedo of the stand is then

$$\text{albedo} = r\,t_d\,T + a\,(1 - T\,(1 - r\,(1 - t_d)))$$

The first term represents reflection from the snow surface and the second term represents reflection from the canopy.

This theory was tested on clear days for three different stands, two with snow on the ground and one without snow, using silicon solar cells as radiation sensors (Federer 1971). Agreement was very good (Fig. 1) except at very large solar zenith angles where difficulties in estimating F may lead to errors both in the model and in correcting the cosine response error of the silicon cells. The relative absorptions of canopy and snow surface depend strongly on solar zenith angle, while total absorptivity changes relatively little.

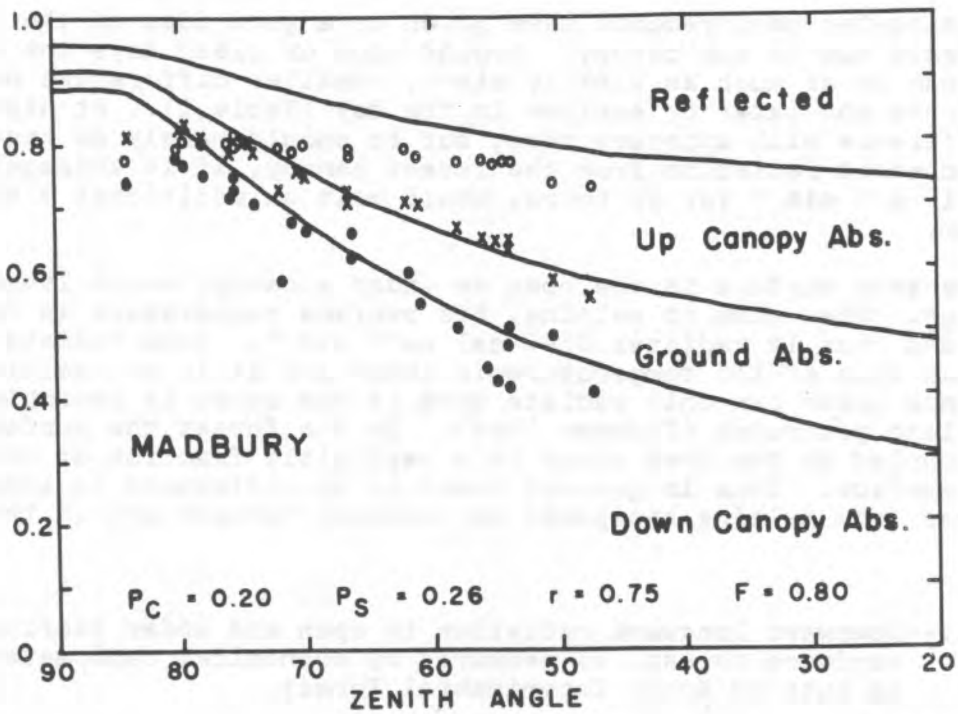


Figure 1.--Reflection and absorption of solar radiation by a hardwood forest in Madbury, N. H. Solid curves are from the theoretical model, points are measured values. From the top of the graph downward, circles are albedo, crosses are reflection from the snow, and dots are transmissivity.

The model still needs extending to include integration over daily periods and to include non-horizontal surfaces. Then we will have a tool for predicting solar radiation regimes for any uniform leafless hardwood forest. Additional work should develop the model to include the radiation regime of openings of various shapes and sizes.

Solar energy absorbed by the canopy may still contribute to snowmelt. Longwave radiation transfers some of this energy from the canopy to the snow surface. And the air in the canopy is warmed so that convective transfer from canopy to snow may also occur.

Longwave Radiation in Leafless Hardwood Forests

The downward flux of longwave radiation above the canopy is determined by atmospheric temperature and humidity conditions, including clouds. Below the leafless hardwood canopy, some of this radiation from the sky is replaced by radiation from the canopy. Normally in melting situations, the effective radiating temperature of the canopy is higher than that of the sky it obscures. The downward longwave flux is thus greater below the canopy than it is above the canopy.

Radiometer measurements have given us a good idea of the amount of increase due to the canopy. Around noon on clear days the increase can be as much as 0.08 ly min^{-1}. Smaller differences occur with clouds and later or earlier in the day (Table 1). At night the difference will approach zero, but it should rarely be negative. This increased radiation from the forest canopy, if it averages 0.04 cal cm^{-2} min^{-1} for 12 hours, would melt an additional 3.6 mm of water.

The snow surface in the open or under a canopy emits longwave radiation. When snow is melting, its surface temperature is fixed at 0°C and thus it radiates 0.46 cal cm^{-2} min^{-1}. Snow radiates less than this if its temperature is lower and it is not melting. But a snow cover can only radiate more if the cover is incomplete or if slash protrudes (Federer 1968). In the forest the surface area occupied by the tree stems is a negligible fraction of the ground surface. Thus in general there is no difference in longwave loss from melting snowpacks in hardwood forests and in the open.

Table 1.--Downward longwave radiation in open and under leafless hardwood forest, as measured by economical radiometers on Hubbard Brook Experimental Forest.

Date	Time, EST	Open: cal cm^{-2} min^{-1}	Below canopy: cal cm^{-2} min^{-1}	Cloud cover
3/9/66	1300	0.32	0.40	0
	1500	.37	.42	3/4
3/22/66	1200	.42	.46	3/4
	1400	.43	.47	1
	1600	.46	.48	1
3/25/68	1400	.32	.37	0
	1600	.32	.36	0
3/26/68	1100	.33	.41	0
	1500	.41	.46	3/4
	1700	.45	.47	1
3/27/68	1300	.40	.44	1/4
	1500	.48	.50	1
	1700	.46	.48	1
3/28/68	1000	.39	.44	0
	1200	.42	.47	1/4
	1400	.49	.50	3/4
	1600	.41	.47	1/2

Armed with bits of knowledge from the literature, from modelling, and from field measurements, we have hypothesized possible radiation budgets for noon on a clear day in late March (Table 2). Downward solar radiation above the canopy is set at 1.00 cal cm^{-2} min^{-1} for convenience. Snow surface albedo is 0.50, corresponding to fairly old snow. Slash protruding above the snow decreases albedo and increases surface temperature (Federer 1968). Net radiation above the canopy of both hardwoods and conifers is much higher than above a plain snow surface and is slightly higher than above snow in moderate slash. However, below the canopy, net radiation in conifers is one-quarter and in hardwoods one-half of that over snow in the open. The effect of canopy absorption of solar radiation dominates the net radiation term and is only partly offset by increased downward longwave radiation. From consideration solely of radiation budgets we can see why snow melts most rapidly in the open, more slowly in hardwoods, and slowest under conifers.

Table 2.--Possible radiation budgets under clear sky at noon in late March. R_{su} is upward or reflected solar radiation, R_{lu} is longwave radiation emitted from the snow surface, and R_n is net radiation.

Location	R_{sd}	R_{su}	R_{ld}	R_{lu}	R_n
	cal cm^{-2} min^{-1}				
Above conifers	1.00	0.10	0.40	0.55	0.75
Above hardwoods	1.00	.20	.40	.50	.70
Open, snow + slash	1.00	.25	.40	.49	.66
Open, snow	1.00	.50	.40	.46	.44
Under hardwoods	.55	.28	.46	.46	.27
Under conifers	.10	.05	.52	.46	.11

Turbulent Transfer

The turbulent transfer processes--convection and condensation-evaporation--are much more difficult to study than is radiation. Energy transfer by these processes is proportional to the temperature difference between surface and air for convection and to the difference of vapor concentration between snow and air for condensation-evaporation. Any time the air is warmer than 0°C, heat is being transferred to the snow by convection and may be used in melting.

When the air has a specific humidity greater than 3.8×10^{-3}, condensation occurs on the snow surface and the latent heat released may be used to melt snow. However, if sufficient energy is not available to maintain melting, hoar frost forms. If, on the other hand, the snow surface has a higher specific humidity than the air, evaporation occurs, cooling the snowpack by latent heat removal. This process is most important in midwinter. The water lost is partly made up by condensation during the snowmelt period.

Net evaporation losses from snowpacks seldom exceed 25 mm a year in temperate North America (Doty and Johnston 1969; West 1959; Williams 1958). As meteorological variables change, five conditions can occur at the snow surface: hoar frost formation, sublimation, evaporation and melting, condensation and melting, and condensation and freezing (Hofmann 1963).

The proportionality factor between the heat flux and the gradient of either temperature or humidity is a variable, and therein lies the major problem of snowmelt prediction. The factor is primarily a function of wind speed, u. A linear relation with zero intercept is often assumed, so the proportionality factor becomes the product of u, several constants (L, C_p, ρ), and a dimensionless turbulence parameter, K. For plain snow surfaces, K is often considered to be a constant (U. S. Army 1956), but this is certainly an oversimplification. From micrometeorological theory for a uniform, horizontal surface

$$K = 0.16 / (\ln \frac{z}{z_o} - \Psi)^2$$

where z is the height at which u, T_a, and q_a are measured, z_o is the roughness length of the surface, and Ψ is a stability correction. The roughness length accounts for turbulent transfer due to mechanical or frictional mixing, and the stability correction allows for turbulent transfer due to buoyant mixing. The roughness length may vary from 0.1 to 0.6 cm for plain snow (Lowry 1969), depending on its surface structure. So if z is 2 m, K can vary from 0.021 to 0.028 when Ψ is zero.

When convective heat transfer, also called sensible heat flux, is zero, the density of the air is constant with height. Then there is no buoyancy effect, and Ψ is zero. When the surface is warmer than the air above, the lower warmer air rises, and the colder air above sinks. This causes turbulence, and Ψ becomes positive. Over melting snow where the surface is colder than the air above, the coldest air is already next to the ground. The atmosphere is stable, turbulence is hindered, and Ψ is negative. Therefore when convective melt is occurring, K is smaller than--perhaps only half of--the value expected for $\Psi = 0$. The stability correction varies with windspeed, roughness length, and sensible heat flux. Theoretical equations for its determination are still a problem in micrometeorology even under ideal conditions.

Beneath a leafless hardwood canopy or in a conifer stand, turbulent transfer theory that works for flow above a uniform surface may no longer hold. In sub-canopy flow, mechanical turbulence is produced over a range of heights, rather than only at the bottom of the atmosphere. In addition, every branch, twig, or needle acts as a heat source; so buoyant mixing also develops from a heat source distributed through the lower levels of the atmosphere. The stability correction may be positive above the canopy and negative below the canopy. In addition, the sources of sensible heat are distributed vertically through the canopy while the sink for latent heat (water vapor) is only at the ground. This means that the turbulence parameters for convection and for condensation-evaporation may not be equal. In short, we do not know what values to use for K when we are concerned with snow under plant canopies.

We made some measurements at Hubbard Brook to get some idea of differences in convection melt and condensation melt between a hardwood forest and an open area (a devegetated watershed). On 4 days in late March 1968, we found little difference between the two areas in air temperature and humidity at 1 1/2 m height (Table 3). With air temperature between 10 and 12°C, the open was less than a degree warmer than the hardwood forest. If wind conditions were similar in the two areas, this temperature effect indicates that convection melt in hardwoods would be at least 95 percent of that in the open. Differences in air humidity were even smaller. We can conclude that any differences in convection and condensation melt between hardwoods and the open are not due to differences in air temperature and humidity. They can be due only to differences in turbulence parameters and wind speed.

Table 3.--Mean temperatures and humidities for hardwood forest and open (devegetated watershed) at Hubbard Brook. Data from two locations in hardwoods and two in open.

Date	Time interval	Number of measurements		Temperature °C		Specific humidity x 10^{-3}	
		Open	Hdwd	Open	Hdwd	Open	Hdwd
3-25	1220-1400	3	2	0.4	0.6	1.9	2.2
	1400-1700	3	4	1.1	1.9	1.1	1.5
3-26	1040-1200	3	3	6.6	6.8	2.0	2.2
	1400-1630	5	5	7.6	7.6	2.6	2.4
3-27	1200-1400	6	6	11.7	11.4	5.1	5.0
	1430-1700	8	9	11.2	11.1	5.3	5.0
3-28	0830-1105	8	8	9.6	9.2	4.5	4.1
	1105-1335	5	7	12.3	11.6	4.8	4.9
	1335-1510	8	6	11.7	10.7	4.4	4.6
	1510-1730	7	9	10.8	10.5	4.6	4.4
Mean	--	56	59	9.4	9.3	4.1	4.0

There is no doubt that the hardwood canopy reduces wind speed. We also measured wind in the hardwood forest and in the devegetated watershed. Anemometers were mounted at 1 m and 5 m heights at the same times and same four locations for which we obtained air temperature and humidity data. On all four days the midday winds were from the south, upslope. We were surprised to find that winds in the forest were consistently about 1 m sec^{-1} or more and were one-half to two-thirds the speed of the wind in the open (Fig. 2). Furthermore the winds in the woods seemed more steady, and the two woods locations agreed better than did the two locations in the open. Lack of agreement in the open is due to the upper elevation being on a meso-relief ridge and 50 m higher in elevation than the lower location, which is in a meso-valley. During the day, wind speed was greater at the 5 m height on each tower than at 1 m height. But as wind in the open died in the evening, nocturnal cold air drainage began and wind speed at 1 m in the forest began to exceed wind at 5 m (Fig. 3). This height reversal was not observed in the open. Evidently wind speed in a leafless hardwood forest may often be 50 percent of that in the open. If we make the unproved assumption that the turbulence parameters in hardwoods and in the open are equal, then convection-condensation melt in the forest is about half of that in the open.

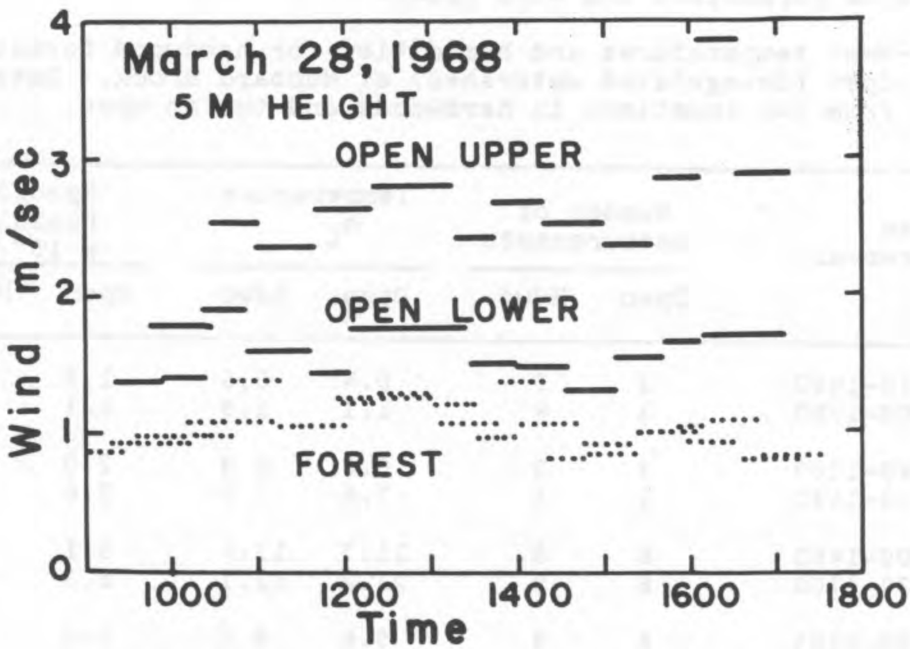
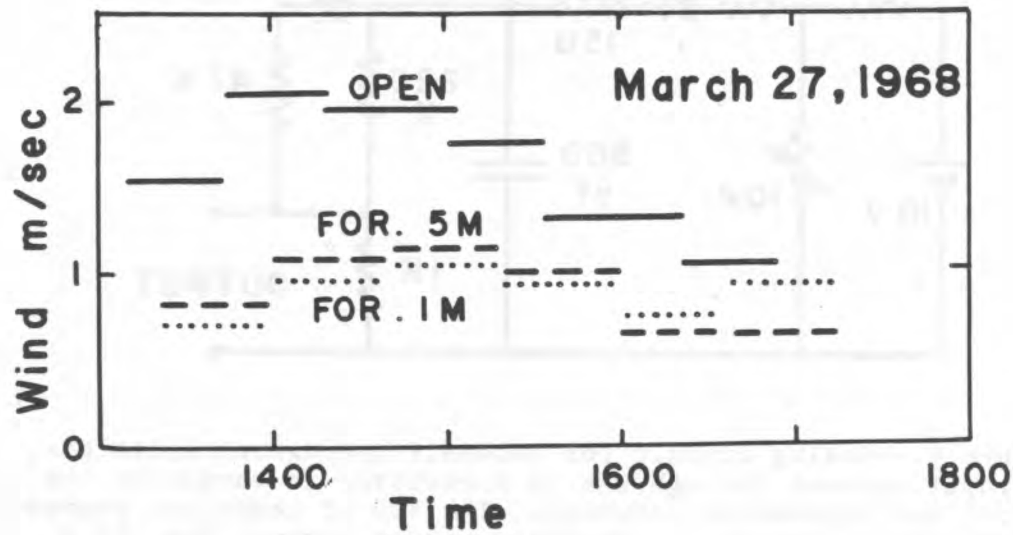


Figure 2.--Wind speed at two locations in the open and two in adjacent hardwood forest on 28 March 1968. The upper location in the open is 50 m higher in elevation than the lower location.

Soil Water Movement and Analog Modelling

So far we have discussed the factors influencing the snowmelt rate. We can reasonably assume that transmission time through and storage in the snowpack are of minor significance. Therefore the rate at which water reaches the soil surface is equal to the melt rate. Yet this water still has a long way to go to become streamflow. Normally in hardwood forest it moves into and through the soil. If rain or extremely high melt rates occur, or if concrete frost or other impeding layers are present, some water may flow on the soil surface. However at Hubbard Brook we may assume that most--if not all--of the flow moves to the stream channel via the soil.

Figure 3.--Wind speed at two heights on the upper forest tower and at 5 m on the upper tower in the open for March 27, 1968.



In the Northeast, late fall rain and midwinter snowmelt is almost always sufficient to eliminate soil water deficits. The spring snowmelt season begins with a soil at or above field capacity. Additional water does not pass all the way through the soil, bypassing water that was already there. New water at the "top" of the soil causes old water to start draining from the bottom. So initial streamflow response can be very fast. Yet there is a delay or storage effect. The soil can get wetter, up to saturation. On the steep soils at Hubbard Brook, where most of the year there is no saturated zone, we find water tables rising toward the surface when the water input rate is high enough. After input stops, then drainage continues for a long time, but at an increasingly slower rate.

Several approaches are useful for studying the relationship of snowmelt to streamflow. Direct measurement of soil-water content and movement leads to understanding the mechanisms involved, but is very difficult on forested watersheds. Measurement of snowmelt at index plots (Davar 1970) is useful for checking snowmelt predictions and as input to a black-box relation between snowmelt and streamflow. Where snowmelt has not been measured directly, we thought a simple electrical analog might simulate watershed behavior well enough that snowmelt could be estimated from streamflow (Fig. 4). We present our analog model here because it seems like a good approach others might want to work on. It looks reasonable, but we have not had time to test it. Measured snowmelt data would be necessary for testing.

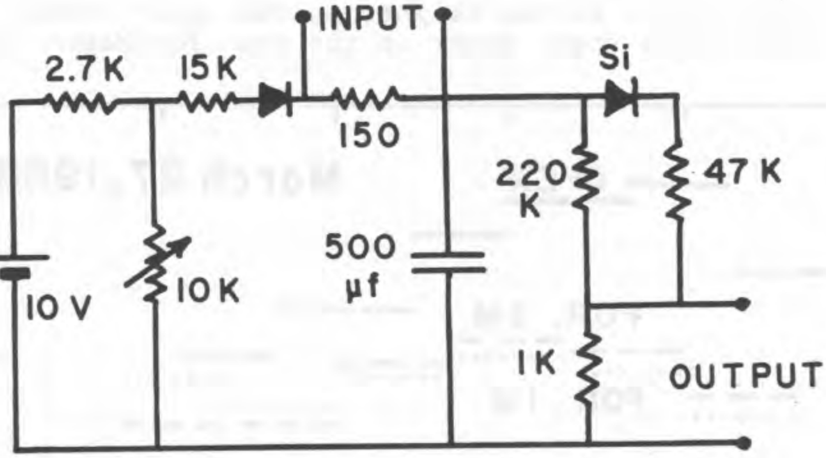


Figure 4.--Analog circuit for snowmelt hydrograph synthesis. An input current through the 15 K resistor is varied by the 10 K pot and represents snowmelt. The 500 µf capacitor represents storage in the soil. Drainage occurs through the 220 K resistor at all voltages (storages) and through the silicon diode and 47 K resistor at higher voltages.

A capacitor drained by a fixed resistor produces an exponential decay or recession curve. At Hubbard Brook this simple recession does not fit measured streamflow except in summer, when transpiration is occurring. In late fall and spring, recession is too fast at high flows and too prolonged at low flows for a simple exponential to fit. So in our circuit we added a smaller resistor in parallel that only operated at higher flows, or higher voltage, because of a diode in series with it. Recession from this circuit fitted measured recessions quite well; so we assumed the model was adequate.

Mechanically, the model operated with two potentiometric recorders, one on the input, and one on the output. Prior to an analog run, measured streamflow was traced manually onto the chart of the output recorder. Then, both recorders running simultaneously, the variable resistor was altered manually to keep the output pen following the output trace. The input pen thus recorded the input necessary to give the required output. Results for a period of mixed snowmelt and rain seemed to be very reasonable (Fig. 5). We know the model can be improved in some respects, notably by the use of a constant current source for input.

Snowmelt in a Stripcut Hardwood Forest

So far all our discussion has been about uniform forest cover or large open areas. Yet we know that small openings in the forest influence snow accumulation and melt. Openings of the right size accumulate snow because of wind behavior and elimination of interception. Snow disappears soonest on the north edge of openings and remains longest on the south edge. These are primarily effects of radiation transfer, but cold air drainage also plays a role. Yet we do not know the significance of this behavior with respect to streamflow. Nor have we quantitatively related melt differences in openings to the meteorological factors causing them.

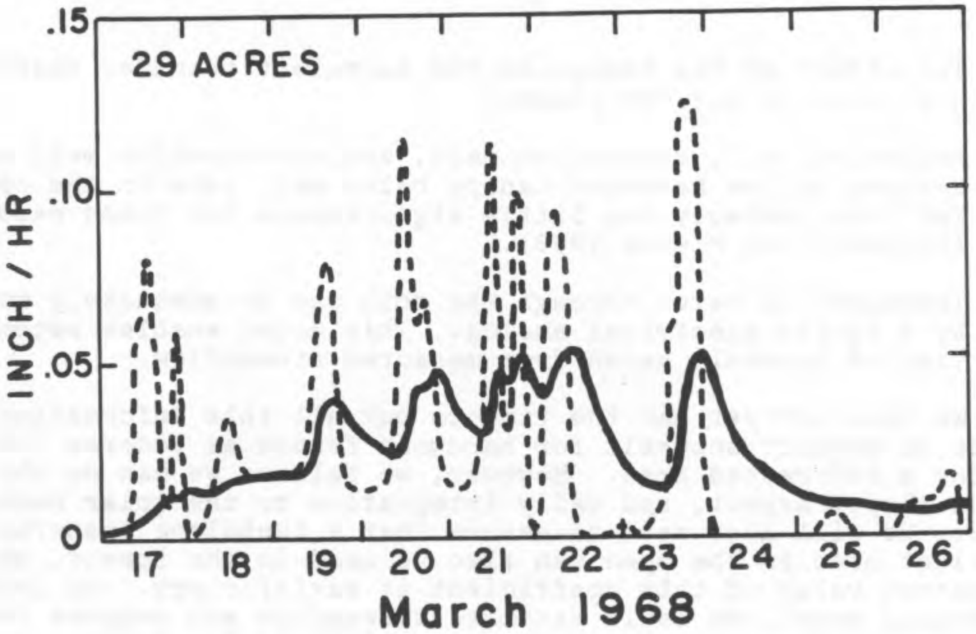


Figure 5.--Input to analog circuit (dashed) required to give an output (solid) equal to measured streamflow from watershed 1 at Hubbard Brook.

Last November watershed 4 at Hubbard Brook underwent the first stage of a three-stage stripcut. The watershed was divided into 25 m-wide east-west strips. Every third one was cut. In 1972 the adjacent strips to the south will be cut, and in 1974 the third and last set of strips will be cut. This offers us an ideal opportunity to study the influence of such strips on snow accumulation and melt in a hardwood forest. At the conclusion of the study we hope to be able to conclude whether or not production of openings in the forest has any effect on snowmelt floods.

In Conclusion

From our work on snowmelt in hardwood forests we have found that, when snow covers the ground:

- The relative amounts of solar radiation absorbed by the canopy and by the snow surface are very variable, depending on solar elevations, sky conditions, slope, aspect, snowpack reflectivity, and canopy structure. However, these can be theoretically modelled.
- During the day, downward longwave radiation is 0.02 to 0.08 cal cm^{-2} higher below the canopy than above.
- Air temperature and humidity are practically the same under hardwoods and in nearby large open areas having the same slope and aspect.
- Wind speed is reduced by only one-half to two-thirds by the presence of the canopy.

- The effect of the canopy on the turbulent transfer coefficients over snow is not yet known.
- Radiation melt, convection melt, and condensation melt are all reduced by the hardwood canopy below melt rate in the open. Yet this probably has little significance for flood reduction (Hornbeck and Pierce 1969).
- Transport of water through the soil can be adequately modelled by a simple electrical analog. This model enables reconstruction of snowmelt rates from measured streamflow.

We have not yet had the time to put all this information together to predict snowmelt for hardwood forest as Federer (1968) did for a deforested area. However, we believe we can do this after we add slope, aspect, and daily integration to the solar radiation model. We will also need to assume that a turbulent transfer coefficient used in the open can also be used in the forest, and that a constant value of this coefficient is satisfactory. By including the analog model, we could estimate streamflow and compare the estimate with measured streamflow from Hubbard Brook watersheds.

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