SNOW IN DIFFERENT ROUGHNESS ZONES AT SCHEFFERVILLE:

ITS CHARACTER AND HYDROLOGIC SIGNIFICANCE

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ABSTRACT The Knob Lake drainage basin is subdivided into different roughness zones according to vegetation and topography. The variability in snow depth, density and water equivalent is investigated. The analysis shows that when boundaries are excluded the average water equivalent is largely independent of roughness zone. Boundaries accumulate greater amounts of snow. The average snow density is inversely related to vegetation roughness which gives an increase in average depth with increasing vegetation roughness. The variability in snow depth, density and water equivalent increases with decreasing vegetation roughness and with increasing topographic roughness. The variability in water equivalent greatly influences the length of the snowmelt period.

INTRODUCTION

Wind has a profound influence on the snowcover in the Schefferville area and elsewhere over large parts of the Canadian North. The effects of the wind are modified by the roughness of the ground surface which results in important differences between the snowcovers accumulated in different roughness zones.

This paper discusses the relationship between surface roughness and snow accumulation, and examines the effects of different roughness zones on the snowcover in the Knob Lake (IHD) basin. The paper also discusses how the snowcover functions as water storage and how this is affected by the environment in which the snow is deposited.

RELATIONSHIPS BETWEEN SURFACE ROUGHNESS AND SNOW CONDITIONS

Snow conditions are related to surface roughness through the influence of surface roughness on the near-surface wind. This wind has three main effects. First, it redistributes the snow causing spatial variations in depth and water equivalent. Secondly, transport by wind causes fragmentation of snow crystals, giving variations in the initial snow density. Thirdly, wind affects latent and sensible heat transfers influencing snow metamorphism and promoting the formation of crusts.

Snow Transport

Snow transport takes place approximately according to Melnik's formula (Komarov, 1954):

$$Q = cU^3 \qquad (1)$$

where Q is the amount of snow transported in the lowest 2 m of the atmosphere (gcm^{-1}), c is a proportionality factor and U is the wind velocity at standard height ($msec^{-1}$). According to this formula, if U changes with a factor of two, the drift transport changes with a factor of eight.

Snow particles travel by surface creep, or sliding and rolling of snow particles along the surface; by saltation, where the particles move forwards in a series of short jumps; and by suspension, where particles are lifted higher into the airstream by turbulent eddies (Bagnold, 1939). Saltation is always the prevailing mode of transport, regardless of windspeed. According to investigations by Komarov (1954), Shiotani and Arai (1967) and Oura (1967), about 90 per cent of total transport in the lowest 2 m of the airstream takes place within 20 cm of the snow surface. Thus the average distance travelled in each displacement is short and the snow transport responds very quickly to changes in the near-surface wind. Particles in suspension are carried for longer distances between impacts with the snow surface, and the response to a wind change is, therefore, slower. According to Budd, Dingle and Radok (1966), the transport of snow in suspension shows an exponential decline with height. There is also a general decrease in particle size with height (Budd, 1966).

Effects of Surface Roughness on Snow Accumulation

Surface roughness elements may be divided into (i) topographic roughness which has a height to width ratio considerably smaller than unity, and (ii) vegetation roughness which generally has a height to width ratio considerably greater than unity. The two roughness types influence the airstream in different ways. Topographic roughness induces fixed spatial variations in wind speed causing the wind speed to increase near the surface of convex elements and decrease in concave parts of the terrain. These variations cause inversely related variations in snow accumulation since erosion prevails in zones of acceleration and deposition prevails in zones of deceleration. In flat terrain, there may be variations in wind speed as a result of turbulence in the airstream but these variations are not fixed spatially. They may give temporary surface drift formations but do not produce permanent variations in snow accumulation.

Vegetation roughness reaches into the airstream so that much of the wind energy is dissipated above the ground surface. The overall near-surface wind speed and amount of drift transport are, therefore, reduced in the presence of trees. This effect increases with forest density. Spatially fixed variations of the near-surface wind are greatest when the density of the forest is low. In open woodland, tall, wide trees behave like large buildings in a city, tending to increase the wind speed locally and create wind-scoops around the trees. On the other hand, small trees slow down the airflow near them and for some distance downwind, creating tail-drifts. As forest density increases, the wind-scoops and tail-drifts become less pronounced because of a general lowering of the near-surface windspeed. In open areas, brush vegetation and other minor vegetation roughness have a similar influence on airflow near the surface, though on a small scale. The importance of this type of roughness, however, diminishes through winter as it is wholly or partly buried by snow accumulation. The effects of roughness changes through winter have been investigated in some detail by the author (Granberg 1972). The investigation showed that snow accumulation in open areas gives a progressive reduction in surface roughness. This leads to a corresponding increase in the amount of snow drifting and to an increase in the distance travelled by individual snow particles. This explains, in part, the changes in snow density through winter.

The near-surface wind adjusts rapidly when it crosses a roughness boundary into a new roughness zone. For example, if snow carrying wind enters a forest edge from a lake, erosion ceases while deposition continues as long as there is snow in the air. Since the transport in the lowest 20 cm of the airstream is adjusted to changes in wind speed within a few tens of metres, the greatest effect of the boundary occurs immediately behind the forest edge, where a zone of deep accumulation is observed. This zone contains most of the snow that was moved towards the forest edge by surface creep and saltation, and a good part of the snow in suspension. The outfall of snow in suspension takes place at a rate determined by (i) the variation in snow concentration with height, (ii) the terminal fall velocity of the snow at the different levels and (iii) the wind velocity profile. Since snow concentration and fall velocity decrease while wind velocity increases with height, not only does the total amount of snow in the air decrease rapidly away from the boundary but the remaining snow is also spread over a larger and larger area. Thus, while blowing snow may be carried for long distances in over a forest, the additional water equivalent deposited dwindles rapidly with distance from the forest edge and soon becomes insignificant. Analysis of sequence melt photographs (Granberg and Thom, 1970; Granberg, 1972, 1973a; Nicholson, in this volume) indicates that in the Knob Lake basin the water equivalent added by blowing snow becomes insignificant at about 100 m from the boundary. This distance varies somewhat depending on the character of the boundary and is shorter for boundaries of greater roughness and longer for boundaries of lesser roughness. The size of an open area and its topographic roughness are important factors in determining the vertical distribution of snow in an airstream and thus in determining the distance over which increased accumulation is significant beyond a roughness boundary.

When an airstream leaves a zone of greater roughness and enters a zone of lesser roughness (eg. from a forest to a lake), erosion prevails over deposition for the distance over which the airstream is accelerated and for some distance beyond. The distance needed for snow transport to reach an equilibrium between erosion and deposition (called "the limit length of snow storm" by Dyunin, 1967) varies with snow, surface roughness and wind conditions. This equilibrium is important in determining differences in snow accumulation between vegetation cover types (Findlay, 1966; Adams and Findlay, 1966; Adams et al, 1966). "The limit length of snow storm" according to Dyunin (op cit), varies from 200 to 500 m or more and varies indirectly with the friability of the snow surface. It is also influenced by topographic roughness which induces turbulence in the airstream, affecting the transport of snow in suspension. Snow accumulation in the forest has been reviewed by Jeffrey (1968) and Meiman (1968). In the case of openings in a forest, small openings have a water equivalent greater than the average for the surrounding area. The water equivalent inside the opening decreases to a value smaller than average as the size of the opening increases. The decrease is probably caused by two different processes: (i) drifting into the opening of snow originally intercepted by trees (Gary, 1974) and (ii) wind depletion of the snowcover in the opening. Both are influenced by the relationship between the size of the opening and the length of its boundary. Openings that are small enough to largely prevent wind erosion of the snow surface have a water equivalent greater than snowfall because of snow drifting from surrounding trees. This effect is greatest for small openings and decreases as the ratio of boundary length to area decreases. As the size of the opening increases, erosion of the snow surface begins. The maximum loss should occur for openings approximately the size of the "limit length of snow storm". The average accumulation in the opening again increases with further increase in size as the loss per unit length of boundary becomes constant and this loss is distributed over a progressively larger area.

Density

During transport, repeated impacts with the snow surface cause fragmentation of snow crystals. The fragments pack more closely together than the initial crystals, giving the snowpack a greater density. In Antarctica, Kotlyakov (1961) found a linear relationship between the density of recently deposited snow and wind speed during deposition. Further increase in density, through depth compaction of the deposited snow, is affected by the weight of the snow column. This, in turn, is influenced by redistribution of snow by wind. Thus, density may be seen partly as a function of surface roughness.

Surface crusts

Both latent and sensible heat transfers take place at rates influenced by the steepness of the near-surface wind velocity gradient. This affects snow surface temperatures and near-surface temperature gradients within the snowpack. These gradients are important factors influencing the formation of crusts at the snow surface. Wind variations caused by roughness differences may produce notable spatial variations in surface temperature conditions, eg. the snow surface in the forest can be melting, whilst in the open evaporation occurs at a sufficiently high rate as to keep the snow surface temperature well below freezing. Thus, a melt crust may form in the forest while a wind crust forms in the open, as a result of the different wind regimes.

Evaporation

Evaporation during the period up to peak accumulation does not appear to be important in the Schefferville area. Elsewhere, it has been found (Molchanov, 1960) that evaporation may reduce the snow cover in the forest because of increased absorption of solar radiation by trees causing evaporation of intercepted snow. However, in the Schefferville area, most of the snow is blown off the trees during, or soon after, snowfall. In this case, the trees shade the snow surface rather than enhancing the absorption of solar radiation. During snow drifting, Dyunin (1967) found very high rates of snow evaporation because of the large total snow/air interface when snow is airborne. It therefore seems that, in the Schefferville area, with its high incidence of snow drifting, the potential for snow evaporation is greater in open terrain than in the forest.

SNOW IN THE KNOB LAKE BASIN

Snow surveys of the Knob Lake basin have formed an integral part of the research at the McGill Sub-Arctic Research Laboratory in Schefferville. Since 1965, with a few exceptions, snow surveys of the basin have been made annually, as part of the hydrology program. The early surveys (Findlay, 1966; Adams and Findlay, 1966; Adams <u>et al</u>, 1966; Rogerson, 1967) found marked variations in water equivalent with vegetation cover (table 1). The sampling was carried out at fixed intervals along traverses transsecting the basin, recording the cover type for each sampling point.

A survey by Petch and Price (A.G. Price, pers. comm.) employing a stratified random sampling design, gave a contradictory result showing no statistically significant difference between snow water equivalents in different cover types. In this survey, different cover types were selected and water equivalents sampled within each one.

The Knob Lake Basin Snow Survey in 1969

A basin snow survey was undertaken during the period March 28 - April 4, 1969. The method of sampling was similar to that used by Findlay (1966). Samples were taken at intervals of 100 m along traverses transsecting the basin in different

TABLE 1

Early Surveys Compared to the 1969 Survey
Water Equivalents (Inches)

	N	Mean	Max	Min
	1965[1]			
Open	43	12.3	36.0	3.5
LC	44	11.4	15.3	8.8
LO	57	15.2	30.0	7.5
Lake[2]	65	6.3	17.3	2.5
	1966[1]			
Open	44	13.0	23.8	3.4
LC	22	15.0	20.0	5.1
LO	14	15.5	19.8	13.1
Lake[2]	57	6.2	10.0	2.0
	1969 (Boundary Zones included)			
Open[3]	182	15.7	42.0	0.0
Closed Woodland[4]	18	16.5	22.5	13.5
Open Woodland[5]	37	18.1	30.0	9.5
Lake	–	–	–	–
	1969 (Boundary Zones excluded)			
Open[3]	182	15.7	42.0	0.0
Closed Woodland[4]	15	15.6	17.5	13.5
Open Woodland[5]	24	15.4	19.5	11.0
Lake	–	–	–	–

1. Source: Adams *et al*, 1966.
2. The water equivalent of white ice not included.
3. Recent - Regenerating burn.
4. Equivalent to LC.
5. Equivalent to LO.

directions (Fig. 1). The intervals between the sampling points were determined by pacing and the sampling points were mapped onto aerial photographs. Care was taken to avoid biasing the sample and each sample was taken exactly where it fell regardless of local obstacles. Lakes were not sampled. A Mount Rose sampler was used for the survey. Overestimates of more than 10 per cent have been recorded for this sampler (Beaumont 1967). Tests in the Schefferville area gave overestimates of approximately 8 per cent. Differences between water equivalents sampled by different observers were also noted. In the present survey all samples were taken by the same observer.

Delineation of Roughness Zones

Six different roughness classes were chosen so as to provide a gradation from high to low roughness. The roughness grading was made, firstly, according to tree density and, secondly, according to topographic roughness. This represents a slight modification of the system used by Findlay (1966). The roughness zones thus defined are:

1.	Closed woodland	(tree density greater than 30 per cent)
2.	Open woodland	(tree density 10 - 30 per cent)
3.	Regenerating burn	(tree density less than 10 per cent)
4.	Recent burn	(trees almost totally absent)
5.	Bog	(tree density less than 10 per cent; flat terrain)
6.	Lake	(trees absent; flat terrain)

The roughness zones were delineated using a set of aerial photographs taken when the ground was snow covered. Individual snow sampling points, mapped on the photographs, made it possible to stratify the sample into sub-strata according to roughness zone. A boundary zone of 100 m was imposed on each side of roughness boundaries involving zone 1 and 2 and samples falling within this zone were assigned to a special class. Since lakes were not sampled and only a few samples fell within group 5, the discussion of these two groups in this paper is based on studies made subsequent to the 1969 basin survey.

Patterns of Snow Accumulation

The areal distributions of snow depths, densities and water equivalents are shown in Figure 2. The horizontal axis in the diagrams represents a unit area. The vertical axis represents the magnitude of the individual parameter. The diagrams thus give a direct areal representation of the variation in the different parameters.

Snow cover in closed woodland (Fig. 2a)

Little or no erosion can take place at the snow surface when the density of the forest is greater than 30 per cent. The snow accumulation in open spaces between trees is, therefore, even. Some snow is intercepted by the canopy, but most of it is blown off the branches, at least in the higher parts of the trees. The snow from the branches is distributed in the open spaces. Spruce and fir in closed woodland usually have dead branches near the ground which gives a less pronounced canopy effect. The water equivalent of the snowpack therefore increases rapidly with distance from the tree trunk. The average density is low and shows little variation since the snow crystals remain intact, unaffected by drifting. The average water equivalent, as determined from 15 samples, is 15.63 inches (397 mm)

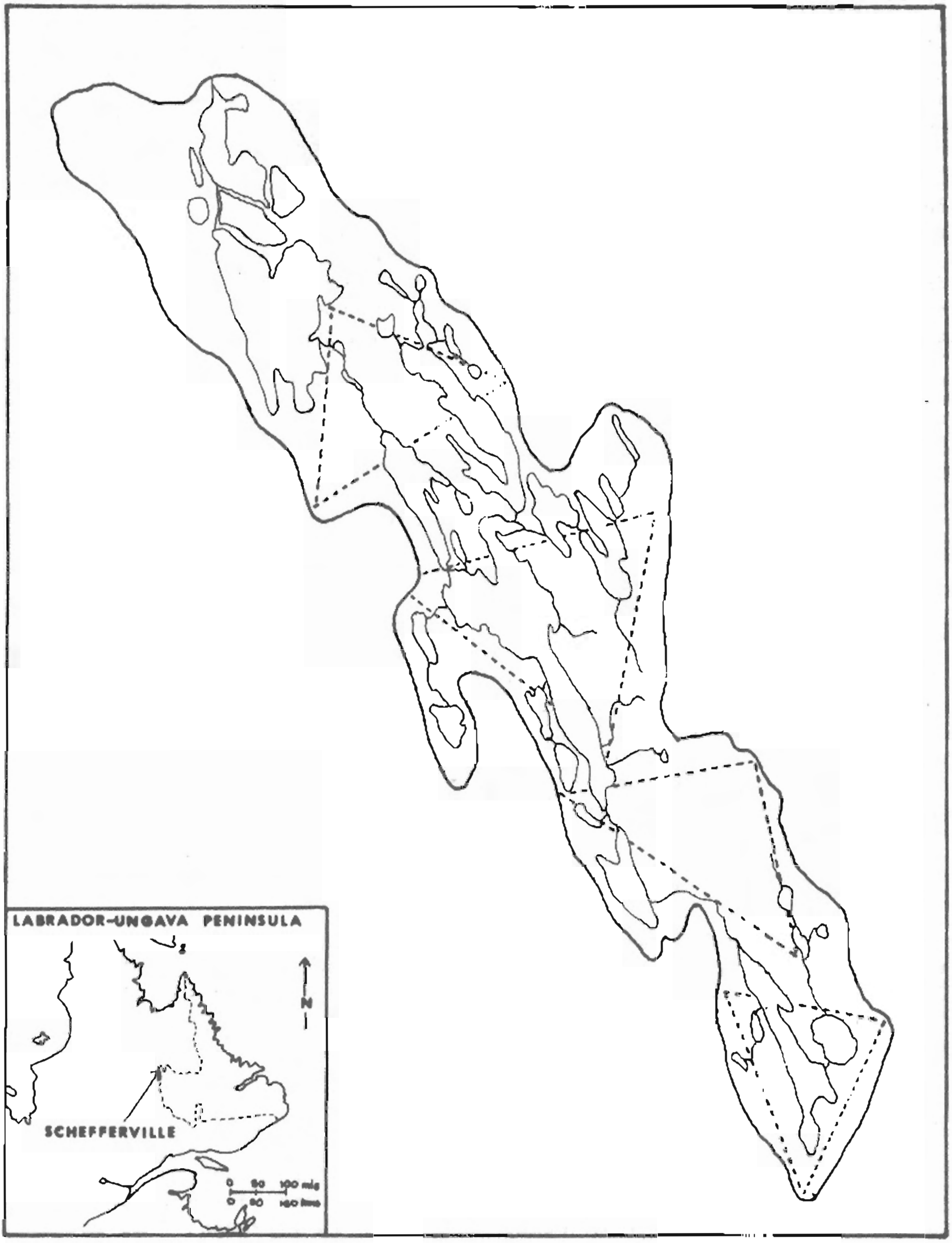
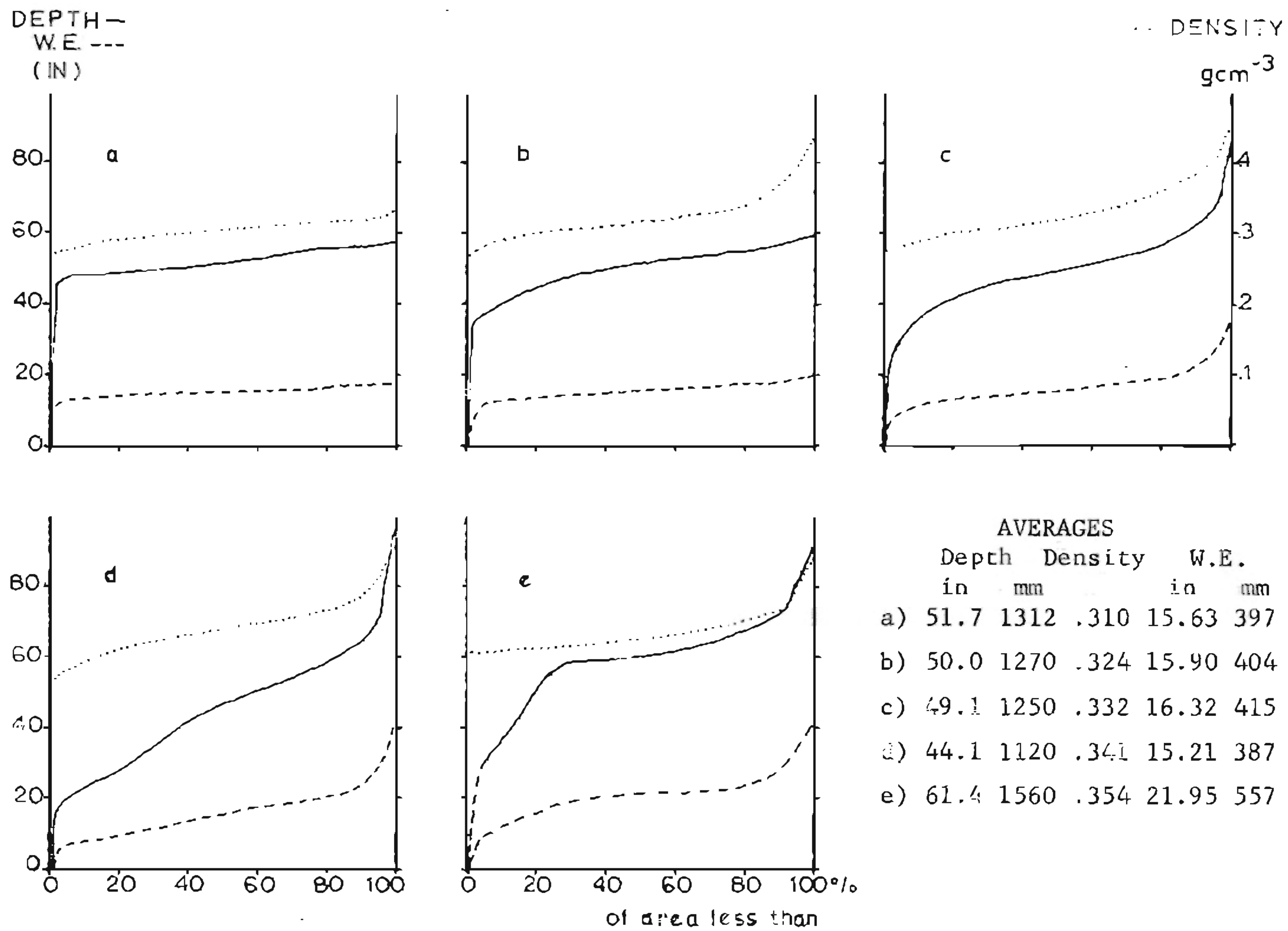


Figure 1. Sampling traverses, Knob Lake basin, 1969



AVERAGES

	Depth in	Depth mm	Density	W.E. in	W.E. mm
a)	51.7	1312	.310	15.63	397
b)	50.0	1270	.324	15.90	404
c)	49.1	1250	.332	16.32	415
d)	44.1	1120	.341	15.21	387
e)	61.4	1560	.354	21.95	557

Figure 2. Depth, densities and water equivalents in (a) closed woodland, (b) open woodland, (c) regenerating burn, (d) recent burn and (e) roughness boundaries.

which is only .07 of an inch (2 mm) less than the overall basin average of 15.70 inches (399 mm). The maximum measured water equivalent in closed woodland was 18 inches (457 mm) and the minimum 13 inches (330 mm). The minimum water equivalent is always zero because of the spaces occupied by tree trunks. These spaces, however, normally occupy less than one per cent of the area and none of the 15 samples actually hit a tree trunk.

Open woodland (Fig. 2b)

Because there is a limited amount of drifting, the snowcover in open woodland shows a greater variability in depth, density and water equivalent than it does in closed woodland. In open woodland, living branches usually extend down to the ground, producing a greater canopy effect than in closed woodland. In addition, there is some variation in water equivalent caused by drift transport along the snow surface. Near the ground surface, wind velocity is increased close to trees and in narrow passages between trees and thus snow is eroded. The wind slows down in the wider openings between the trees where the drifting snow is deposited. This gives a general increase in water equivalent with distance from trees with the greatest rate of increase within a zone 0 - 1 m outside the edges of the trees (Granberg, in press). The average water equivalent, determined from 24 samples, is 15.90 inches (404 mm) which is .20 of an inch (5 mm) greater than the overall average. The snow density is slightly greater than for closed woodland and shows a greater variation.

Regenerating burn (Fig. 2c)

The regenerating burn is characterized by moderate to strong redistribution of snow. The trees are young and of relatively small height and width which makes them aerodynamically quite different from the mature trees in open woodland. The young trees create tail drifts during storms. These drifts harden during the intervals between storms and are an important factor in the overall accumulation pattern, producing cross-bedding effects. As a result, instead of an increase in water equivalent with distance from the trees, there is normally a decrease. The drift transport is sufficient to produce considerable differences in accumulation with respect to topography. The maximum water equivalents are usually found in the lee of sharp drops in the terrain and in smaller valleys. Frequently, there is a well established cover of brush vegetation in the regenerating burn, which retains a certain amount of snow. Thus, there are only relatively small areas completely lacking a snowcover or retaining only a shallow cover of snow. The greater drift transport causes a higher average density and a greater density range than in the two previous roughness zones. The average water equivalent, determined from 75 samples, is 16.32 inches (415 mm) which is .62 of an inch (16 mm) greater than the overall average.

Recent burn (Fig. 2d)

In recent burn, the absence of trees allows a generally high wind velocity near the surface promoting strong redistribution of snow. Topography controls the wind patterns and thereby the patterns of snow accumulation. Redistribution of snow by the wind produces a general smoothing of the terrain with deep accumulation in valleys and little or no snow on ridge crests. The distribution of depths and water equivalents is skewed, with a large areal percentage covered with shallow snow and a small areal percentage exhibiting deep accumulation. The average density is greater than in the three previous roughness zones but the range in density is only slightly greater. The average water equivalent, determined from 107 samples, is 15.21 inches (387 mm) which is .49 of an inch (12 mm) below the overall average.

Bog

Within the Knob Lake basin this roughness zone is similar to regenerating burn with respect to tree coverage. In contrast, though, the terrain is flat. Thus, there is little or no variation in snow accumulation due to topography. One additional factor, influencing snow accumulation in this environment, is the wet nature of the ground. If snowfall occurs before the ground is well frozen, partial melting of this snow takes place. The bottom layer of the snowcover, therefore, frequently consists of frozen slush. When the snowcover is sampled using snow tubes, this frozen slush layer is usually not accounted for and may give the erroneous impression that less snow accumulates on bogs than in other roughness environments. If the bog is small, drift transport from the bog into surrounding forest causes a depletion of the water equivalent but for bogs of large size this effect is of insignificant magnitude.

Lake

Lakes have the lowest roughness, allowing a greater total drift transport than occurs in any of the other roughness zones. As a result, the snowcover on lakes has the highest average density. As with bogs, the lakes cause partial melting of the snowcover. Snow falling onto a lake before freeze-up is usually completely melted while snow falling later during winter is partially converted into white ice by water penetration onto the ice through cracks (Jones, 1969). This makes it extremely difficult to obtain a good estimate of the water equivalent of the snowcover. The depth of snow on top of the ice varies through winter but is usually less than half the depth of snow in forested country. The amount of snow incorporated into white ice was estimated by Findlay (1966) to be in the order of one third of the water equivalent of the white ice. Subsequent tests have indicated a figure slightly above 40 per cent.

Roughness boundaries (Fig. 2e)

This group includes only boundary zones involving closed and open woodland. It does not include boundaries such as those between recent and regenerating burn. The average water equivalent, determined from 16 samples, is 21.95 inches (557 mm) which is 6.25 inches (159 mm) greater than the overall average. The snow density variations are relatively similar to those of open woodland but the overall density is slightly greater.

Differences in Snow Water Equivalent

The present survey shows that snow water equivalents in the different cover types differ only slightly (see table in figure 2). The greatest difference is between regenerating burn and recent burn, but the mean of those two is only .05 of an inch (1 mm) below the overall average. These results thus conform with the observations of Petch and Price, but disagree with the results of the early surveys (Table 1) which showed marked differences between accumulation in different cover types. The inclusion of boundary zones in the 1969 survey (also Table 1) produces differences between cover types although not as pronounced as the early surveys. The suggestion is that the main difference between cover types is caused by accumulation in boundary zones.

Although the drift transport on lakes is greater than any other roughness zone, it is reasonable to assume that the average loss of snow from lakes in the Knob Lake basin is normally not in excess of one to two inches of water equivalent.

The loss should be greater for lakes bounded by forest and smaller for lakes bounded by low roughness zones.

According to values given by Adams et al (1966), and subsequent surveys by the present author, variations in water equivalent of the snowcover on lakes are considerably smaller than those for open terrain. This indicates that the variability in water equivalent increases with increasing topographic roughness, but also decreases with increasing vegetation roughness as is shown by Figure 2.

Density and Depth Variations

Both the average density and the variability in density increase with decreasing vegetation roughness (Fig. 2). The average depth decreases with decreasing vegetation roughness, as a result of this difference in average density.

THE WATER STORED AND RELEASED BY THE SNOWCOVER

The hydrologic role of the snowcover is affected by the way in which it accumulates. While total water equivalent is the main factor influencing the amount of water released during spring melt, the variability in water equivalent is a major factor determining the timing of the snowmelt water release, because it determines the rate of reduction in snow surface area through melt.

Snowmelt Water yields from different roughness zones

The daily melt water yield, W, from the snowpack may be written:

$$W = Ma \tag{2}$$

where M is the melt rate for the day and a is the average snow covered area for the particular day. In the case of spatially uniform melt rates, the decrease in snow covered area is directly controlled by the spatial variations in water equivalent. This is valid even with some variation in melt rates as long as there is not a strong spatial correlation between melt rates and water equivalents.

The percentage cover was computed for two dates, May 30 and June 6, 1972, and the results compared with observed snow coverage measured on air photos taken on those dates. Transformations of the distributions obtained in the 1969 basin snow survey were used in the computation. To account for the different average water equivalents, 1969 to 1972, a percentual transformation was made by setting the averages for the two years at 100 per cent. The percentage cover was calculated by subtracting the observed melt (Table 2). Average melt was determined from 10

TABLE 2

Predicted and Observed Snowcover in Per Cent of Total Area

Date	Closed Woodland Pred.	Closed Woodland Obs.	Open Woodland Pred.	Open Woodland Obs.	Regenerating Burn Pred.	Regenerating Burn Obs.	Recent Burn Pred.	Recent Burn Obs.	Total Melt (in.)
May 30	100.0	100.0	100.0	100.0	98.0	93.2	95.0	77.2	5.2
June 6	83.0	41.0	81.0	30.9	70.5	77.2	57.0	54.0	11.2

snowcourses in the basin (data from J. Fitzgibbon, pers. comm.). No account was taken of the relatively small differences in melt rates between different zones. The percentage snowcover was measured from the air photos by counting of grid intersections. Despite the assumptions involved there is a good general agreement between the results obtained from the two different methods except for zone 4 on May 30 and zone 1 on June 6.

The water yield from different roughness zones was computed from the melt rates (Fig. 3). The figure shows that, in the early part of the melt, there is little variation in water yield. As the melt rate increases the water yield in zone 1 reaches a peak while the cover is near 100 per cent. Water yield then decreases rapidly as the snowcover is reduced to zero over a short period of time. Although the disappearance of the snowcover in zones 1 and 2 may in reality be somewhat prolonged because of microscale variations in melt rates caused by tree shadows, analysis of air photographs taken during melt in the Schefferville area shows that zones 1 and 2 are completely cleared of snow well before zones 3 and 4. In zone 4 the maximum water yield occurs earlier than in zone 1 but the maximum yield is only about 70-80 per cent of that in zone 1. The melt period is prolonged considerably in zone 4. The same is true for roughness boundary zones where snow remains after zones 1 and 2 are completely snow free.

The average snow water equivalent, the variability in water equivalent and the timing of the onset of rapid snowmelt in spring determine the rate of release of melt water and in particular determine whether high melt rates will coincide with total or near total snowcover and for how long the two will coincide. This is most important for snowmelt flood forcasting. The results given above indicate that forested terrain in the Schefferville area gives a high water yield over a short period of time while snow in open terrain releases its water more gradually. Observations at the Timmins 4 Permafrost Experimental Site, a tundra site 15 miles northwest of Schefferville, shows that the greater range in water equivalent in years of heavy snowfall (Granberg, 1973b) prolongs the snowmelt period quite considerably in those years. Coincidence of an unbroken snowcover with a high melt rate is more likely in years of heavy snowfall than in years of little snow when the snowcover becomes discontinuous before very high rates of melt occur.

The equation for determining the average daily meltwater yield for the basin may be written:

$$W_B = PW_1 + PW_2 + PW_3 + PW_4 + PW_5 + PW_7 \qquad (3)$$

where W_B is the total water yield from the basin snowcover, P is the percentage of each cover type and W_1, W_2 ... are daily melt water yields calculated for each cover type using equation (2). Cover type 7 represents roughness boundaries. Cover type 6, lakes, has been left out of the equation intentionally for reasons which are discussed below.

Hydrologic effects of the snowcover on lakes

Hydrologically, the snowcover on lakes is distinct from the snowcover in other roughness zones, since it does not store water if the outlet of the lake is open throughout the winter. Instead, snowfall raises the hydrostatic water level of the lake and displaces water from the lake in direct proportion to its total water equivalent (providing the outlet geometry remains unchanged). This assumes that the ice cover may be regarded as free floating. Judging from crack patterns observed in spring melt photographs, only a narrow zone along the shore does not comply with this assumption.

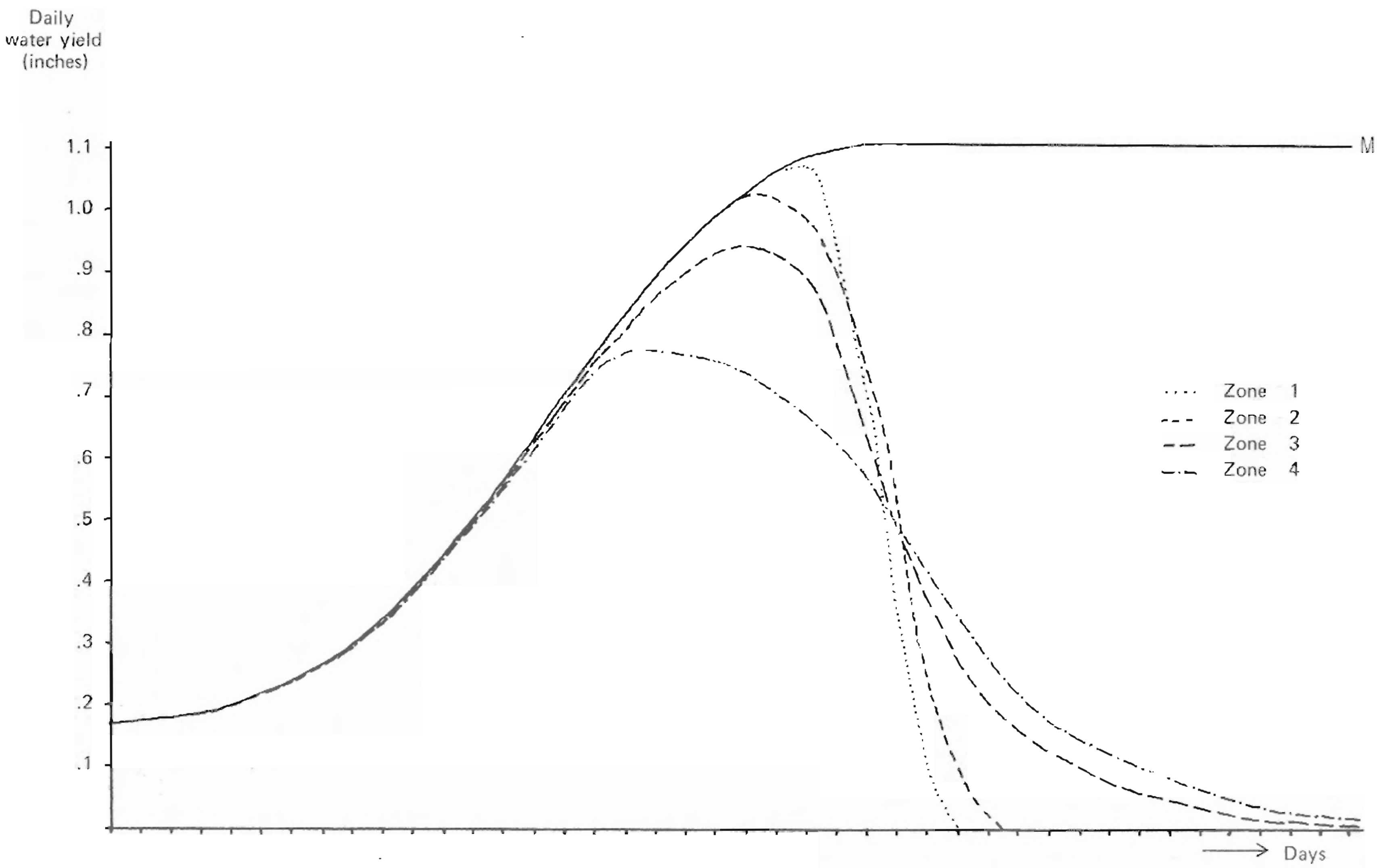


Figure 3. Variations in snowmelt water yield through an idealized snowmelt season as a function of the variability in snow water equivalent in (1) closed woodland, (2) open woodland (3) regenerating burn and (4) recent burn. M is the assumed daily melt rate, equal for all zones.

Displacement of water has been observed in the Knob Lake basin during early winter by Findlay (1966, p. 59; 1967, p. 122, figs. 1-2) who described coincidence of snowfall, strong winds, drifting snow, increase in total ice thickness and increase in the thickness of white ice. Jones (1969) suggests that this displacement is caused by the freezing of ice. However, ice formation, whether it involves white ice or ordinary ice, cannot cause displacement if the ice cover is floating free, since it does not produce any change in the mass balance of the lake, nor in its hydrostatic water level. An increase in water level can occur only if there is a positive change in the mass balance. In winter, when no melt occurs, this can be caused only by snowfall onto the lake or by changes in the geometry of the outlet or inlet stream channels. Unfortunately, the use of stop-logs at the Knob Lake outlet produces a somewhat artificial hydrograph (Findlay, 1966; Penn, 1970). This explains the absence of hydrograph response from December on in the hydrographs shown by Findlay (1966) and Penn (1970). Findlay's hydrograph was obtained at the outlet of Knob Lake and Penn's hydrograph was obtained at the outlet of Pearce Lake. In both cases, the smoothing effect of the stop-logs and the town water supply is sufficient to obliterate flow peaks caused by snowfall on the lakes. Unfortunately, the only available comparable record is from the Menihek power station, 30 miles south of Schefferville, which is subject to the same constraint. The displacement of water caused by the snowcover opens the possibility for a new means of measuring snowfall. A large lake, where boundary effects are small in relation to total lake surface area, could be used as a snow pillow by monitoring its mass balance through the winter.

Lakes are not included in equation (2) since snow melt on lakes does not contribute to spring runoff. Even if the outlet channel has been closed throughout winter, the rate of runoff does not relate to the rate of melt on the lake. Instead, the runoff occurs as a surge as soon as the outlet channel is opened. Observations of surges have been made in the Knob Lake basin at Easel and Houston lakes by John Fitzgibbon (pers. comm.).

SUMMARY AND CONCLUSIONS

This paper has shown that snowcovers accumulated in different roughness zones exhibit systematic differences with respect to average density and depth and with respect to the variability in water equivalent, depth and density. Average water equivalent is largely independent of roughness zone except for boundary zones. This agrees with the results obtained by Petch and Price. A possible exception is lakes, where losses in the order of 1-2 inches water equivalent may occur. The disagreement with results from early snow surveys appears to be at least partially due to the inclusion, in these cases, of boundary zones in computing the averages for the different vegetation zones.

The study indicates that important improvements of spring runoff predictions can be made by separate consideration of the water equivalent distributions in different roughness environments. It also outlines the peculiar hydrological role played by snow accumulation on lakes. This aspect has previously been neglected by research in the Knob Lake basin.

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