

## Controls of Canopy Structure on Snowmelt Rates in the Boreal Forest

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### ABSTRACT

Information on spatial variations in snowmelt rates in the boreal forest is required for hydrological studies conducted as part of the BOREAS field experiment. Several factors may control the spatial variability of snowmelt, including slope angle, slope aspect, and vegetation cover. However, vegetation properties probably exert the greatest control on spatial variations in snowmelt in basins with limited topographic variability. Links between snowmelt and canopy structure are examined for a low-relief drainage basin in the boreal forest near Thompson, Manitoba. Micrometeorological measurements were made at two towers located in open and forested sites, respectively, and were used to compute daily snowmelt by the energy balance approach. Daily melt was also measured using snow-wires at the towers and at four other sites in the basin with differing canopy density characteristics. Information on canopy structure was obtained by measuring the gap fraction of the canopy at each snow-wire site. The ability of an energy balance approach to estimate melt was assessed through a comparison with measured melt. The energy balance provided a reasonable estimate of melt in the open fen; however, overestimation of net radiation at the forest tower resulted in systematic overpredictions of melt. There was good agreement between snow-wire estimates of daily melt and the corresponding canopy structure characteristics as derived from the gap fraction measurements, with open sites experiencing more rapid melt than sites with a closed canopy. The relationship between canopy gap fraction and melt rate suggests that point melt data could be distributed in the basin based on ground-based or remotely-sensed measures of canopy density.

Keywords: snowmelt, boreal forest, canopy structure

### INTRODUCTION

The **BO**REAL Ecosystem-Atmosphere Study (**BO**REAS) is a large-scale field experiment that seeks to elucidate the interactions of energy and mass between the boreal forest biome and the lower atmosphere in order to clarify their roles in global change (Sellers *et al.* 1993). One of BOREAS' objectives is the development and validation of water balance models for the boreal forest. These models can be used in turn to test remote sensing algorithms which facilitate scaling-up micro-scale hydrological properties and processes to calculate regional and global-scale fields of water exchange between the atmosphere and the boreal forest (Sellers *et al.* 1993). These tasks require information on spatial variations in snowmelt rates at various scales, since snowmelt is the dominant hydrological event in the boreal forest. Previous work has shown that a variety of factors may control snow accumulation and melt within a basin, including terrain type (e.g. Woo and Marsh 1978), slope angle and aspect (e.g. Hendrick *et al.* 1971, Price and Dunne 1976, Buttle and McDonnell 1987), and vegetation cover (e.g. Adams 1976, FitzGibbon and Dunne 1983, Burkard *et al.* 1991). However, the limited topographic variability of much of the boreal forest landscape suggests that spatial variations in vegetation properties likely exert a relatively greater control upon melt rates. We hypothesize that melt rates are directly related to gap fraction (the fraction of the sky visible to a sensor located beneath a canopy), and that gap fractions can be used as the spatial fabric to distribute melt rates within the boreal forest landscape. Increased canopy density reduces both incoming

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shortwave radiation ( $K\downarrow$ ) to the snowpack as well as wind speed over the snow, resulting in lower turbulent fluxes of sensible ( $Q_H$ ) and latent ( $Q_E$ ) heat. These reductions in energy contributions are assumed to outweigh increased longwave radiative fluxes ( $L\downarrow$ ) from the canopy to the snowpack as a result of greater canopy density, leading to lower melt rates with decreased canopy gap fraction. This paper examines the influence of vegetation canopy structure on snowmelt in a low-relief environment by addressing the following questions: (1) what is the spatial variability of melt observed within vegetation units?; (2) does melt vary between vegetation units?; and (3) are any variations in snowmelt explained by differences in canopy characteristics?

## STUDY AREA AND METHODS

The study was conducted in a small ( $\sim 2 \text{ km}^2$ ) low-relief drainage basin in the boreal forest 40 km northwest of Thompson, Manitoba ( $55^\circ 55.5'N$ ,  $98^\circ 25'W$ , Fig. 1). The basin discharges to a larger fen which is a BOREAS tower flux measurement site, and consists of low-angle slopes draining to a chain of interconnected wetlands. The gently rolling terrain is underlain by discontinuous permafrost, extends from 253 m to 276 m a.s.l., and is covered primarily by black spruce (*Picea mariana*), with lesser amounts of jack pine (*Pinus banksiana*), paper birch (*Betula papyrifera*) and tamarack (*Larix laricina*). Ground cover on the forested slopes consists of Labrador tea (*Ledum groenlandicum*) underlain by feather moss (*Hylocomium* spp.) and lichen (*Cladina* spp.). Willow (*Salix* spp.) comprises part of the understory but is more prominent at wetland fringes. Characteristic

wetland species include sedges (*Carex* spp.), mosses (*Sphagnum* spp.) and bog laurel (*Kalmia* spp.). The fen to the south of the basin has a sparse overstory of bog birch (*Betula glandifera*), tamarack and black spruce, and an understory of buckbean (*Menyanthes trifoliata*), sedges and cottongrass (*Eriophorum angustifolium*). Wetlands contain typic and humic mesisols, while the hillslopes are mantled by orthic grey luvisols and orthic lutric brunisols. Mean annual precipitation for the basin is  $\sim 0.45 \text{ m}$ , mean annual snowfall is  $\sim 1.7 \text{ m}$ , mean January daily temperature is  $\sim -25^\circ\text{C}$  and mean July daily temperature is  $\sim 16^\circ\text{C}$  (Hydrological Atlas of Canada 1978).

Spatial distribution of snow water equivalent (SWE) within and between vegetative units in the basin was determined prior to melt from a snow course consisting of 62 sampling sites. Since variability of snow depth generally exceeds that of snow density, 5 snow depths and 1 snow density measurement were taken at each site. Depth was measured at one central location and four sites 2 m away in the cardinal directions. Snowpack densities were measured from vertical snow cores obtained using an Atmospheric Environment Service snow sampler. SWE in the fen was measured using a snow course of 95 sampling sites (Bryant 1994, personal communication). Snow depth and density were recorded at each site.

Daily snowmelt was measured using the snow-wire technique (Heron and Woo 1978, Fig. 2). A taut horizontal wire 1.75 - 2.0 meters in length was suspended over the snowpack surface. The distance from the wire to the snowpack surface ( $z$ ) was measured at 10 centimeter intervals over a distance of 1 m along the central portion of the wire prior to the start of the daily melt cycle. Three samples of the surface snow density ( $\rho_s$ ) were obtained at the same

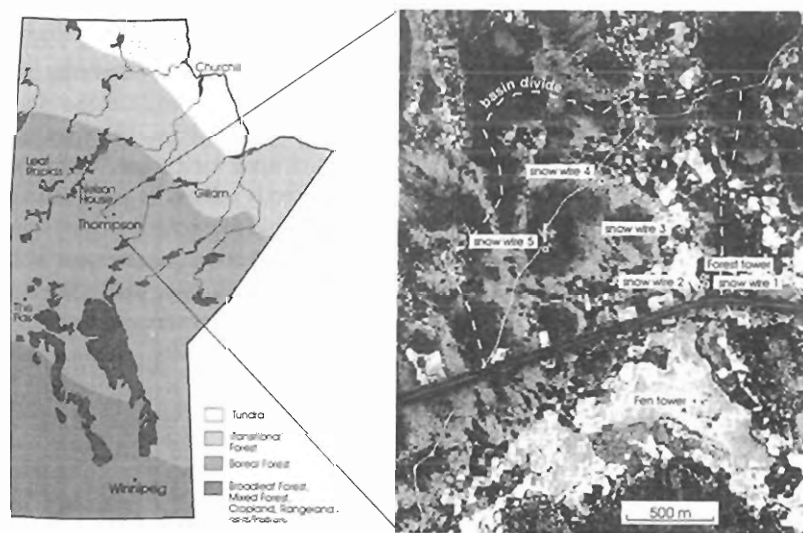


Figure 1. Study basin location and instrumentation (air photo taken on Julian Day 110).

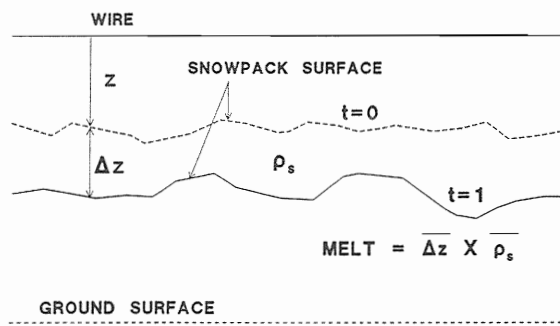


Figure 2. The snow-wire method of estimating melt.

time using 250 cm<sup>3</sup> snow cores extracted parallel to the surface. Melt was determined from:

$$M = \overline{\Delta z} \times \overline{\rho_s} \quad (1)$$

where  $M$  = daily melt (mm)  
 $\overline{\Delta z}$  = mean change in the distance from the wire to the snowpack surface between times  $t$  and  $t + \Delta t$  (m)  
 $\overline{\rho_s}$  = mean density of surface snow layer at time  $t$  (kg m<sup>-3</sup>)

The standard error associated with the mean melt depth was determined as:

$$s.e._M = \sqrt{s.e._{\Delta z}^2 \times \overline{\rho_s}^2 + s.e._{\rho_s}^2 \times \overline{\Delta z}^2} \quad (2)$$

where  $s.e._M$  = standard error of the mean melt depth (mm)  
 $s.e._{\rho_s}$  = standard error of the mean snowpack density (kg m<sup>-3</sup>)  
 $s.e._{\Delta z}$  = standard error of  $\Delta z$  (m)

Five snow-wires were deployed in the dominant vegetation units in the basin prior to melt, while a sixth was installed in a large fen to the south of the basin (Fig. 1, Table 1).

Measured melt depths were compared to estimates obtained from solution of the snowpack's energy balance:

$$Q_M = Q^* + Q_H + Q_E + Q_G + Q_P - Q_0 \quad (3)$$

where  $Q_M$  = energy available for melt (W m<sup>-2</sup>)  
 $Q^*$  = net radiation (W m<sup>-2</sup>)  
 $Q_H$  = sensible heat flux (W m<sup>-2</sup>)  
 $Q_E$  = latent heat flux (W m<sup>-2</sup>)

Table 1. Canopy and pre-melt mean snowpack characteristics ( $\pm 1$  s.d.) for dominant vegetation types. SWE means with the same letter are not significantly different between vegetation types at  $\alpha = 0.05$  using Tukey's multiple means comparison.

Vegetation type	% of basin	Snow-wire	Gap fraction	Depth (m)	Density (kg m <sup>-3</sup> )	SWE (m)
wetland with overstory	-	fen	0.971	0.62 $\pm 0.08$ n = 95	151 $\pm 27$ n = 95	0.094 <sup>A</sup> $\pm 0.018$
wetlands - open	10	2	0.992	0.64 $\pm 0.1$ n = 60	193 $\pm 31$ n = 12	0.122 <sup>BC</sup> $\pm 0.014$
mixed forest	10	1	0.268	0.52 $\pm 0.08$ n = 30	207 $\pm 62$ n = 6	0.107 <sup>AB</sup> $\pm 0.034$
black spruce	35	4	0.063	0.54 $\pm 0.1$ n = 85	208 $\pm 38$ n = 17	0.109 <sup>AB</sup> $\pm 0.019$
black spruce - open	35	3 5	0.459 0.659	0.66 $\pm 0.08$ n = 75	219 $\pm 30$ n = 15	0.144 <sup>C</sup> $\pm 0.018$
mixed forest - open	10	-	-	0.66 $\pm 0.13$ n = 30	222 $\pm 41$ n = 6	0.144 <sup>C</sup> $\pm 0.024$

- $Q_G$  = ground heat flux ( $W m^{-2}$ )  
 $Q_P$  = energy flux delivered to the snowpack by rainfall ( $W m^{-2}$ )  
 $Q_\theta$  = change in negative heat storage in the snowpack ( $W m^{-2}$ )

Instrumentation was installed on two micrometeorological towers: one in the centre of the fen to the south of the basin, and one in a forest stand in the basin beside snow-wire 1.  $Q^*$ , wind speed, air temperature and relative humidity were measured above the snowpack surface at both sites and recorded on Campbell data loggers. Wind direction and  $K\downarrow$  were also measured at the fen tower. Rainfall was recorded at the fen tower and in the wetland near the basin outlet. The bulk transfer approach (Heron and Woo 1978, Price 1988) was used to calculate  $Q_H$  and  $Q_E$ . Rainfall temperature was assumed to equal air temperature in order to estimate  $Q_P$ .  $Q_G$  was assumed to be of minor importance and was ignored.  $Q_\theta$  was approximated by summing nocturnal  $Q^*$  losses from the snow surface prior to the start of positive  $Q^*$  fluxes to the snowpack.  $Q_\theta$  was then satisfied by subsequent  $Q^*$  contributions before the pack was assumed to become isothermal and capable of releasing meltwater.

The LAI-2000 plant canopy analyzer (LI-COR Inc., Lincoln, Nebraska) was used to provide an indirect measure of canopy structural characteristics at each snow-wire. These measurements are presented here in the form of gap fractions which range between 0 (no sky visible to the sensor) and 1 (no foliage visible to the sensor). Gap fractions were determined by using one LAI-2000 unit to record above-canopy sky brightness at 30 s intervals while another unit was used to record below-canopy readings at the snow-wires. Above-canopy readings for snow-wires 1 to 5 were taken on top of a van parked in a clearing at the basin outflow. Above-canopy readings for the fen snow-wire were taken on the roof of a building at the site. Measurements were taken after sunset, before sunrise, or during overcast periods when the sky brightness was closest to optimal (i.e. uniform), in order to minimize the potential error caused by the distance between the two sensors. Heterogeneity of the canopy warranted use of 90° view caps on the sensors; therefore, one measurement was taken in each of the four cardinal directions and averaged to obtain the gap fraction for each snow-wire. The sensors were calibrated prior to measurement, and were aligned in the same cardinal direction during measurement to ensure that both sensors were viewing the same portion of the sky. Gap fraction values for each of the snow-wires are given in Table 1.

## RESULTS AND DISCUSSION

### 1. Spatial variations in premelt SWE

Results from snow surveys on Julian Days (JD) 98, 99 and 100 indicate substantial spatial variability in SWE in the basin prior to melt (Table 1). The fen, black spruce and mixed forest zones had similar mean SWE, which was significantly lower than that in forest areas with more open canopies (open black spruce and open mixed forest). Relatively greater SWE in the more open forest zones likely reflects the influence of airflow perturbations in the open canopy combined with decreased interception and subsequent sublimation of snowfall (Schmidt 1991). Wetlands in the basin had significantly higher peak SWE than the larger, wind-swept fen. Snow accumulation in the fen was limited by wind transport of snow to the adjacent forest, while formation of low density depth hoar at the snowpack base reduced the overall density of the pack. The combination of these factors appears to account for the fen's low mean SWE.

### 2. Limitations of the snow-wire method

Heron and Woo (1978) noted that the snow-wire method can overestimate melt prior to ripening of the snowpack. The snowpack may settle during this ripening phase, which would result in an overestimate of  $\Delta z$ . Density of the surface snowpack layer increased at all sites during ripening (Fig. 3), and melt was determined for the period following this initial rise in  $\rho_s$ . Melt estimates were discontinued towards the end of the melt period, when excessive melt-out near the posts supporting the reference wire made estimates unreliable.

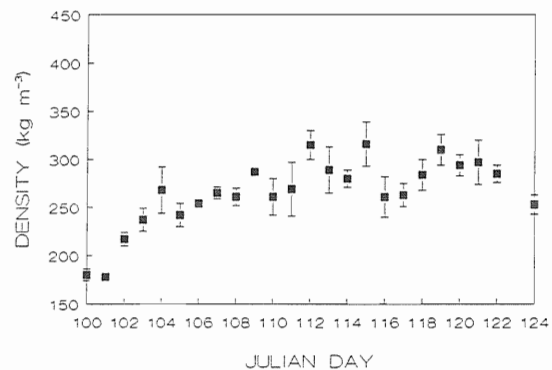


Figure 3. Mean density of the surface snow layer ( $\pm 1$  s.d.) during the study period at snow-wire 1. This pattern was typical of that observed at other sites.

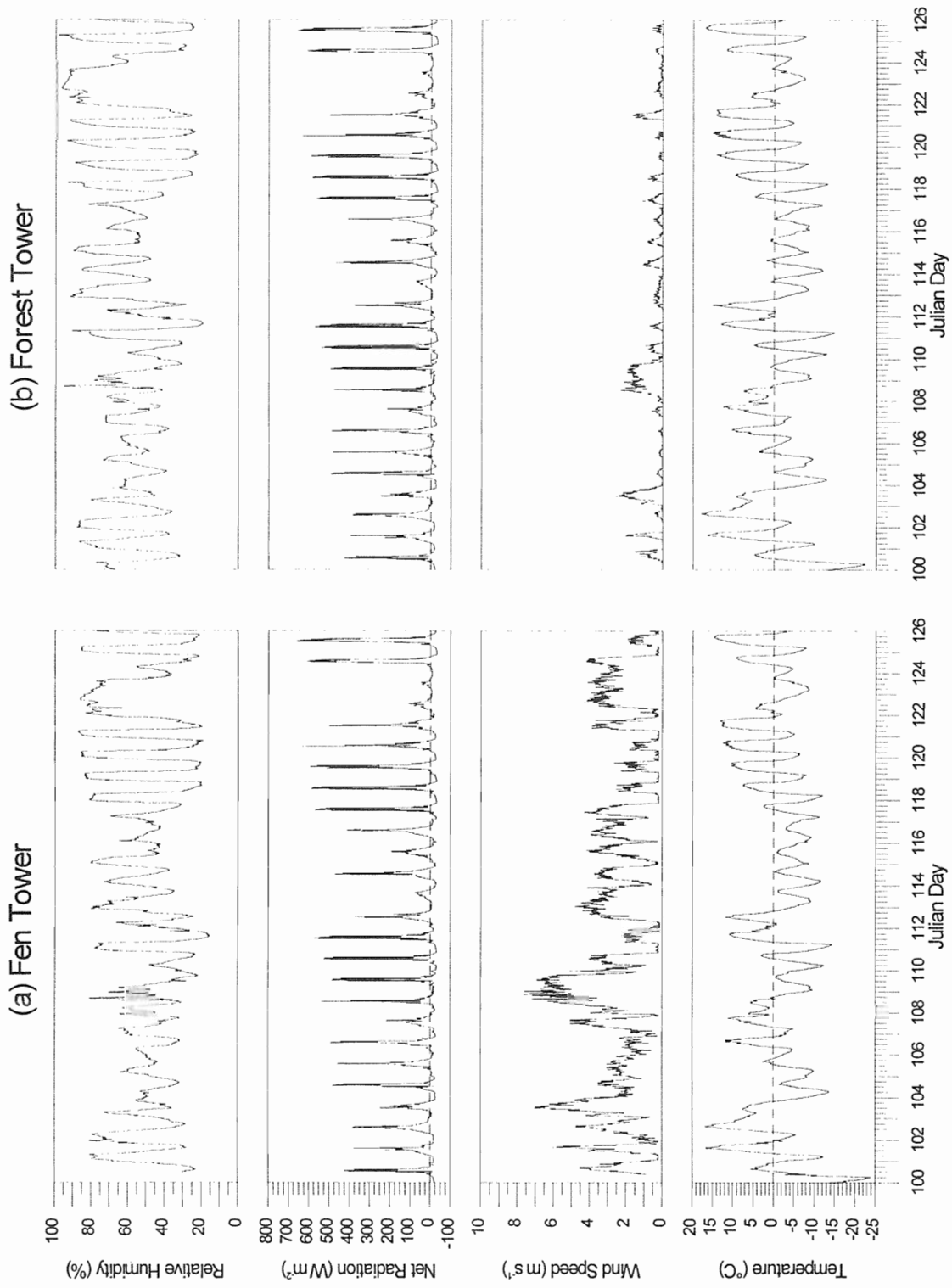


Figure 4. Micrometeorological observations at the fen (a) and forest (b) towers during the 1994 snowmelt period.

### 3. Climatic conditions during melt and energy balance estimates of melt

Snow-wire measurements and observations at both towers began on JD 100 (Fig. 4). Maximum daily air temperatures were above 0°C prior to this date; however, low nocturnal air temperatures (e.g. -24°C at 0500 h on JD 100) delayed warming of the snowpack until JD 100. The snowpack quickly became isothermal by the afternoon of JD 101, resulting in rapid initiation of melt. Melt then became more gradual with greater nocturnal cooling of the pack, ceasing completely between JD 113 and JD 116 when maximum air temperatures remained below 0°C. A warming trend beginning JD 117 quickly warmed the pack and reinitiated melt.

Energy balance estimates of daily melt were determined for the fen and forest towers for the periods JD 102-109 and JD 102-118, respectively. Loss of snowcover beneath instruments at the forest tower prevented calculation of melt after JD 118. A comparison of daily melt measured at the fen snow-wire with estimates from the energy balance measurements at the fen tower (Fig. 5a) shows reasonable agreement similar to that obtained by Heron and Woo (1978) at a high arctic site using the same basic techniques. The best-fit regression relation between observed melt at the fen snow-wire (*SW*) and predicted melt using the energy balance (*EB*) from JD 102-109 was:

$$SW = 2.7 + 0.8EB; R^2 = 0.61; s.e._y = 6.2 \text{ mm} \quad (4)$$

However, the intercept term in eq. 4 is not significantly different from 0, while the slope term is not significantly different from 1. Data from the fen snow-wire suggest enhanced melt near the support posts on JD 109, and removal of this point improves the correspondence between observed and estimated melt. The energy balance provides a better estimate of cumulative melt at the fen snow-wire site (Fig. 5b); however, the more rapid initial melt at snow-wire 2 indicates that there may be substantial error associated with applying the energy balance estimates from the fen tower to other open sites in the basin.

There was no systematic over- or underprediction of observed melt at the fen; conversely, the energy balance method at the forest tower systematically overestimated daily melt measured at snow-wire 1 (Fig. 6). This reflects the dominant control that  $Q^*$  exerted on  $Q_M$  at the forest tower during the 1994 snowmelt, as has been shown in other years at other boreal forest sites (e.g. Price and Dunne 1976, FitzGibbon and Dunne 1983).  $Q_H$  and  $Q_E$  made minor contributions to melt; therefore accurate predictions of

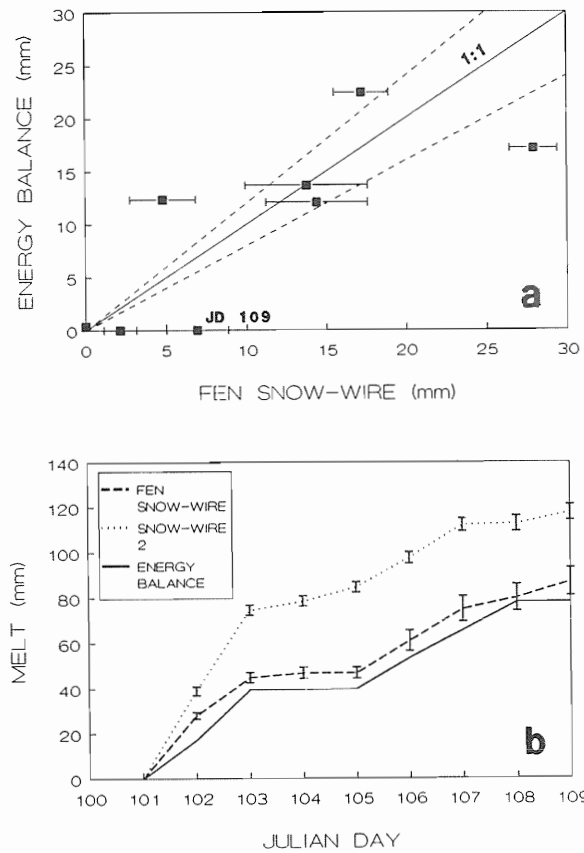


Figure 5. (a) Daily melt measured at the fen snow-wire ( $\pm 1$  standard error about the mean) vs. melt estimated from the energy balance at the fen tower. Dashed lines indicate  $\pm 20\%$  about the 1:1 line; (b) cumulative melt measured at the fen snow-wire and at snow-wire 2 ( $\pm 1$  standard error about the mean), and estimated from energy balance measurements at the fen tower.

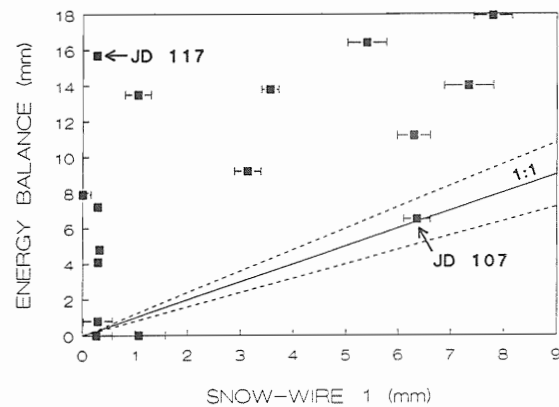


Figure 6. Daily melt measured at snow-wire 1 ( $\pm 1$  standard error about the mean) vs. melt estimated from the energy balance at the forest tower. Dashed lines indicate  $\pm 20\%$  about the 1:1 line.

$Q_M$  depend on obtaining representative estimates of  $Q^*$  over the snowpack surface using a single radiometer ~1.5 m above the snow surface. Shadows cast by the trees mean that at times the upper radiometer dome is shaded, while the underlying dome "sees" a relatively bright surface, resulting in low  $Q^*$  values. At other times, the upper dome receives direct sunlight ("sun flecking") while the lower dome records outgoing short- and longwave radiation from a surface covered in shadows, resulting in pronounced spikes in  $Q^*$ . Nadeau and Granberg (1987) deployed 20 net radiometers in an open boreal woodland. They noted that "sun flecking" resulted in individual instantaneous  $Q^*$  values that exceeded the areal mean  $Q^*$  by >100%. Such spikes might be expected to result in an overestimate of hourly averages of  $Q^*$ , which would in turn be used in eq. (3) to estimate melt. Figure 7a demonstrates the magnitude of this problem for a sunny day, where "sun flecking" resulted in spikes in the measured  $Q^*$ , an overestimate of total radiative energy input to the snowpack, and a substantial overprediction of melt (Fig. 6, JD 117). The problem of "sun flecking" is not present on overcast days (Fig. 7b), with the result that  $Q^*$  more closely follows the diurnal pattern of  $K\downarrow$  and the energy balance method provides a much better estimate of observed melt (Fig. 6, JD 107). These results highlight the importance of obtaining representative values of  $Q^*$  when attempting to model snowmelt in forests using an energy balance approach. These could be obtained by averaging the signals of multiple radiometers installed a smaller distance above the snowpack surface (e.g. Nadeau and Grandberg 1987).

#### 4. Melt estimates from the snow-wire method

Melt measurements at snow-wire 2 and the fen ceased on JD 110 and JD 112, respectively. Measurements at snow-wires 3, 4, and 5 continued until JD 124, while melt at snow-wire 1 ceased on JD 126. However, melt estimates obtained near the end of the melt period are unreliable due to loss of continuous snow cover beneath the snow-wires, and these data have not been included in subsequent analyses. Substantial variability in melt was observed between the various sites (Fig. 8a). As expected, the open sites (fen and snow-wire 2) ablated much faster than the forested sites. However, daily melt could differ significantly between sites located within the same vegetation unit (Fig. 8b). For example, initial melt was much more rapid at snow-wire 2 than at the fen wire (Fig. 8a); following this period, melt depths were comparable at the two sites. Higher melt at snow-wire 2 relative to that at the fen may be attributed to three factors: (1) dust contributions from Highway 391 at the

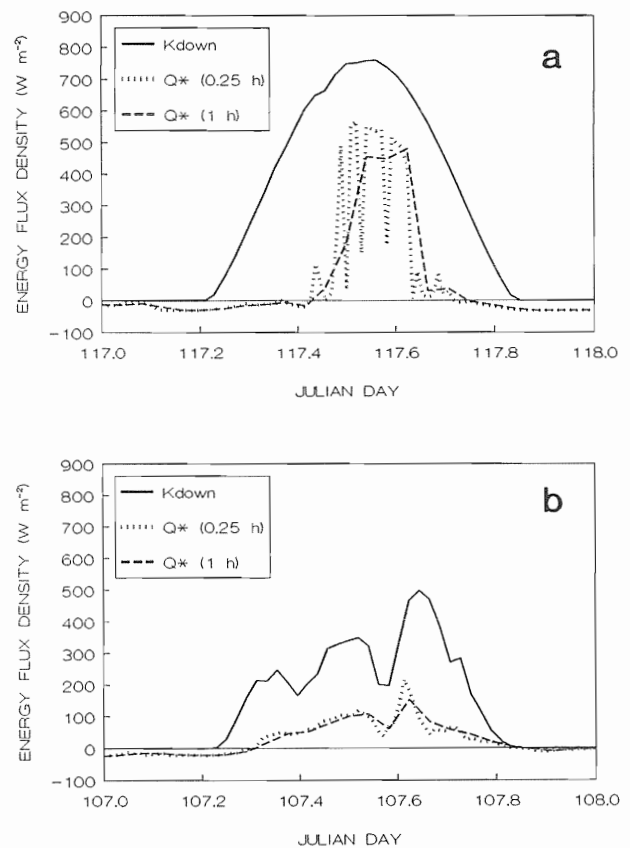


Figure 7. Incoming short-wave radiation ( $K\downarrow$ ) and 15 minute and hourly averages of  $Q^*$  over the snowpack surface at the forest tower: (a) JD 117; (b) JD 107.

southern boundary of the basin, which would reduce the snowpack's albedo close to the road; (2) presence of the sparse overstory at the fen, which would reduce wind speeds and turbulent fluxes relative to the open site at snow-wire 2; and (3) greater advection of sensible heat from the surrounding forest at snow-wire 2 relative to the fen site, which was less susceptible to boundary influences owing to its greater fetch.

Cumulative melt also varied appreciably between the four snow-wires located in the forest (Fig. 8a). This suggests that a simple discretization of the basin into open (wetland) and forested areas would obscure significant spatial variability in melt. Variations in melt pattern can be largely attributed to differences in canopy density, as indicated by the gap fraction values at each snow-wire site (Table 1). Closed-canopy sites (e.g. snow-wire 4) tended to experience slower, more protracted melt than more open canopies (e.g. snow-wire 5), as has been noted in other studies of snowmelt in boreal forests (e.g. FitzGibbon and Dunne 1983). This is supported by Figure 9, which indicates an inverse relationship between gap fraction and the time required to ablate 50 mm of SWE at the snow-wire

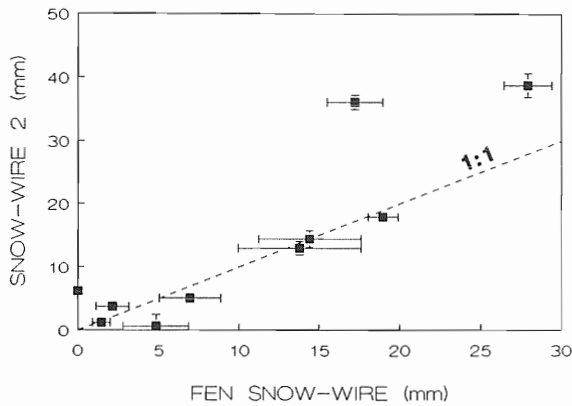
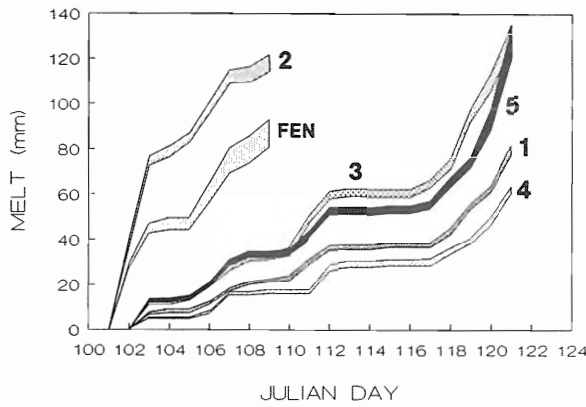


Figure 8. (a) Cumulative melt at the six snow-wire sites. The range in cumulative melt indicates  $\pm 1$  standard error about the mean melt depth; (b) daily mean melt ( $\pm 1$  standard error about the mean), fen snow-wire vs. snow-wire 2.

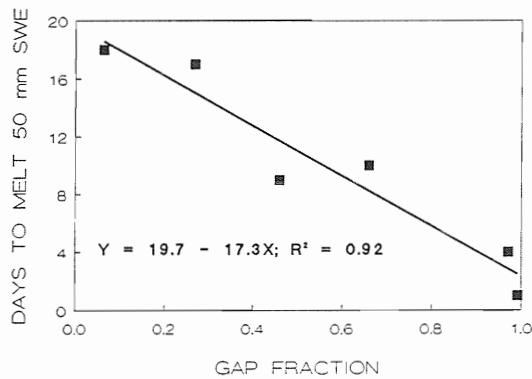


Figure 9. Days required to melt 50 mm of SWE vs. canopy gap fraction measured at the snow-wire sites.

sites. However, canopy density as measured by gap fraction does not provide a complete explanation of spatial variations in melt, and snowcover at snow-wires 3 and 5 ablated at similar rates despite a 58% greater canopy density at the former site. Ground cover at

snow-wire 3 included Labrador tea, which was not present at the other sites. These plants were covered by snow at the start of melt; however, they began to emerge above the snow surface on JD 110-111. This coincides with increased melt at snow-wire 3 relative to snow-wire 5, and Price and Dunne (1976) note that emergence of brush through a thinning snowcover tends to reduce the snow's albedo, leading to enhanced melt. Gap fraction measurements were taken late in the melt period at mean height of 0.4 m above the ground surface. Thus, the Labrador tea understory would have reduced the measured gap fraction at snow-wire 3 relative to the other sites which did not possess this understory. Canopy density during the initial melt at snow-wires 3 and 5 may have been more similar than is suggested by Table 1, in keeping their similar cumulative melt patterns.

## SUMMARY

Snow accumulation and melt were found to vary significantly between vegetation types in a low-relief basin in the boreal forest. An energy balance approach to estimating daily melt performed adequately for a fen site with sparse overstory; however, the energy balance consistently overestimated melt at a mixed forest site with a relatively closed canopy. This appears to reflect the influence of shadows falling on the upper dome of the radiometer and the underlying snowpack surface upon the hourly mean  $Q^*$  estimated using a single instrument. There was good general agreement between spatial variations in melt rates within the basin and the corresponding canopy gap fraction. This suggests that canopy characteristics exert a strong control on melt within the basin, and that basin-wide estimates of snowmelt can be obtained by distributing point melt rates using information on the spatial variation of canopy density. This information could be obtained from ground-based measurements and/or remote sensing platforms, and may provide a physically-meaningful approach to the regionalization of point hydrological processes and properties required as part of BOREAS.

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