1.4 Hydrological Processes In Cold Regions

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ABSTRACT

The objective of this study was to study three processes important to the water and energy cycles of northern environments, namely: snow accumulation and wind transport of snow; ablation of seasonal snowcover; coupled heat and mass transfer in snow and underlying frozen ground, and to develop physically based algorithms that describe these processes using field measurements in boreal, alpine and arctic environments. Progress has been made in understanding and describing many of the processes in a physical manner, in evaluating the process descriptions and in developing operational algorithms for specific processes. Some coupling, and/or comparisons of process algorithms with standard land surface scheme calculations have been demonstrated. The observed multi-scale operation and horizontal interaction of some of these processes means that phenomena operating at very small scales can affect large-scale water and energy balances. The relative success in transposing hydrological process descriptions from one environment to another can be attributed to the strong physical basis of the descriptions.

The following algorithms of cold regions hydrological processes have been developed, examined with respect to their performance and undergone enhancements where appropriate:

- Blowing snow model – sublimation
- Complex terrain blowing snow model – snow accumulation
- Intercepted Snow Accumulation/Unloading/Sublimation – sublimation
- Accumulation and ablation of shallow snow covers of open environments
- Boreal forest snow-cover – ablation
- Infiltration to frozen soils – operational algorithm

Introduction

Objectives

To study three processes important to the water and energy cycles of northern environments, namely:

(1) Snow accumulation and wind transport of snow;
(2) Ablation of seasonal snowcover; and,
(3) Coupled heat and mass transfer in snow and underlying frozen ground, and to develop physically based algorithms that describe these processes using field measurements in boreal, alpine and arctic environments.

Methodology

Field Work

Intensive measurements of snow accumulation, ablation and frozen soil infiltration were made at four locations:

(1) Inuvik, NWT (IN) with Pomeroy, Marsh and Schuepp
(2) Wolf Creek, Yukon (WC) with Pomeroy, Granger and Woo
(3) Waskesiu, Saskatchewan (PA)
(4) Kernen Farm, Saskatchewan (KF)

Snow accumulation measurements concentrated at IN, WC and KF through winter with an eddy correlation and blowing snow system installed at KF and the Trail Valley site (IN), along with intensive snow surveys. The purpose of the IN measurements was to provide an evaluation of blowing snow model performance in an Arctic environment, whilst KF provided an evaluation in a Prairie environment. At WC an eddy correlation system, suspended spruce tree and intensive snow surveys characterized snow interception, sublimation and accumulation in forest and tundra environments. An eddy correlation unit was placed at IN from April through June 1999 to provide reliable surface sensible and latent heat flux measurements from taiga for comparison to airborne flux measurements by Schuepp.

Snow ablation measurements were conducted in open terrain at KF from March to April and forest and tundra terrain at WC from mid April to late May. Ablation measurements consisted of eddy correlation/energy balance systems on level plains, north and south facing slopes and valley bottoms, transect-snow ablation measurements (1 m resolution), snow covered area and occasional tethersonde atmospheric profile measurements. The purpose of KF measurements was to further test open environment snowmelt energetics calculations. The purpose of the WC measurements was to provide areal measurements of melt energetics in complex terrain as an aid to upscaling snow ablation calculations and to provide input for distributed infiltration simulations.

Infiltration to frozen soil measurements were conducted during snowmelt at PA in April and WC in April and May. Gamma twin probe densitometers, soil thermistors and TDR were measured intensively through the melt period for forest and tundra soils in order to better characterize the processes controlling infiltration in these environments. Several readings per day were conducted for specific sites and for a variety of snowpacks, melt rates, soil textures and initial soil moisture content profiles.

Experimental Sites and Data Collection

Sites for data collection were chosen in the prairie (Bad Lake/Saskatoon), boreal forest (Waskesiu) and arctic (Trail Valley) regions of western and northern Canada. All locations experience severe continental climates with winter temperatures ranging from just above 0^{o}C down to approximately -40^{o}C and relatively low annual snowfall. The mean corrected annual snowfall totals (in terms of water equivalent) at the three sites are of the order of Bad Lake/Saskatoon --120 mm, Waskesiu -- 150 mm, and Trail Valley -- 170 mm.

Bad Lake, Saskatchewan: Creighton Watershed, Smith Tributary

The Bad Lake Watershed (51^{o}23’N, 108^{o}26’W, 650 masl.) in south-western Saskatchewan was established as a Research Basin under the International Hydrological Decade Program and instrumented for hydrological research beginning in the late 1960s. The basin is largely cultivated (≈ 70% of the area) in the production of cereal grains, pulse and legumes by dryland farming. Of the remaining area, 20% is rangeland and 10% is native shrub, grass and farmyards. The topography of the basin ranges from poorly drained, level plains to moderately and steeply rolling upland areas, such as the Creighton and Smith Tributaries, which contain deeply incised channels. Its climate is semi-arid and typical of northern grasslands, with dry, cold winters and hot summers. Snowcover on the basin exhibit high seasonal and interannual variability. In high snow years, the snowcover normally forms in November and disappears in April, in years with light snows the snowcover forms and ablates several times over the winter with final ablation in March. During low snow years, the major accumulations are found in waterways and within and surrounding windbreaks due to redistribution of the snowfall by wind. Experiments to measure snow-covered area, small-scale and large-scale albedo and surface energy balance were conducted during snowmelt periods on 1972 and 1973 (O'Neill 1972; O'Neill and Gray 1973). In 1973 and 1974 extensive snow surveys were conducted over various landscape types during mid-winter and during the snowmelt period (Steppuhn and Dyck 1974).

Saskatoon, Saskatchewan: Kernen Farm

The Kernen Farm located just east of the city of Saskatoon, SK (52°N, 107°W, 500 masl) is an experimental research farm operated by the University of Saskatchewan. It is situated on a flat, open, lacustrine plain, which is cropped to cereal grains and pulse crops under dryland farming. The climate of the area is sub-humid and typical of the northern prairie-parkland transition with cold winters and a snowcover from November through March/April. Experiments were conducted in March 1994 and 1996 on fallow and stubble fields on snow accumulation, snowcover depletion, areal albedo, and the energetics of snowmelt (Shook 1995; Shook and Gray 1997).

Waskesiu, Saskatchewan: Beartrap Creek

The Beartrap Creek basin (53.9°N, 106.1°W, 550 masl), is located near Waskesiu, SK within Prince Albert National Park. Sites in this basin were instrumented in 1992 for a series of forest hydrological process experiments for the Canadian GEWEX and Prince Albert Model Forest studies. The region is dominated by mixed-wood southern boreal forest. Its climate is continental boreal with cold winters and snowcover from late October through April (Harding and Pomeroy 1996; Pomeroy and Dion 1996; Pomeroy et al. 1997a).

Experiments were conducted from a tower in a mature jack pine, *Pinus banksia*, stand in the Beartrap Creek. The site is level, has a winter leaf + stem area index of 2.2 m^2/m^2, canopy coverage of 82% and an average tree height of 19 ± 3 m. Over the period of study from November 1994 to April 1995, the low-lying carpet of sphagnum moss and kinnikinnick (bearberry) was covered by snow.

Inuvik, Northwest Territories: Trail Valley Creek

The Trail Valley Creek basin (68°45'N, 133°30'W, 150 masl), located 50 km north of Inuvik, was instrumented for hydrological process and basin modelling experiments in 1991. Currently, these studies form part of the Canadian GEWEX program. The basin has a low arctic climate with snow-cover from September through May. Vegetation is predominately tundra (70%) that is interspersed with large areas of shrub tundra (21%) and small pockets of transitional forest-tundra (1%) (Marsh and Pomeroy 1996). Deep snowdrifts, which accumulate in the lee of abrupt changes in topography, cover about 8% of the basin (Pomeroy et al. 1997b). Experiments were conducted using observations of micrometeorological variables made from a tower located on a tundra plain with excellent open, uniform fetch characteristics.

Meteorological Stations

At Waskesiu, a variety of meteorological instruments were mounted on the 27-m tower in the center of the pine stand (Pomeroy et al. 1997a). Parameters used to run CLASS (radiation, air temperature, humidity, and wind speed) were measured above the forest canopy. At the top of the tower, two "Kipp and Zonen" short-wave radiometers and one "Middleton" net radiometer measured the incoming and reflected short-wave radiation and net radiation respectively. The radiometers were cleaned weekly and often more frequently when snowfall occurred. Air temperature/humidity was measured using a "Vaisala" HMP35CF hygrothermometer and wind speed was measured using a "Weathertronics" cup anemometer. In addition, the ground heat flux was measured using REBS heat flux plates and snow and soil temperatures were measured using "Type-E" thermocouples. Campbell 21x microloggers were used to control instruments and record data.

At Saskatoon and Trail Valley Creek, 3-m towers were erected in large, uniform, open, snow-covered plains. Similar equipment to that used at Waskesiu was deployed along with an eddy correlation system at Saskatoon. This system comprised a "Solent" ultrasonic anemometer and a "Campbell Scientific" Krypton fast hygrometer. Instantaneous measurements of vertical wind speed, air temperature and specific humidity were made at 10 Hz and covariance calculated over 15 minutes. The eddy correlation system and all other equipment were controlled and logged by Campbell 21x microloggers (Shook and Gray 1997).

At Bad Lake, similar radiation equipment was employed along with standard meteorological equipment used in a climatological station operated by the Atmospheric Environment Service, Environment Canada (Gray and Granger 1988). Snow depth and soil temperature were also monitored.

Snowfall, Accumulation and Interception

Snowfall measurements were made with Nipher-shielded AES-style cylinders emptied daily (Bad Lake) or weekly (Waskesiu) or as necessary (Inuvik). The readings were corrected by procedures described by Goodison et al. (1997). At Waskesiu, one-half-hour estimates of snowfall rate were made using a snow particle detector (Brown and Pomeroy 1989), whose counts were given a mass based on weekly snowfall. Snow accumulation was determined from areal snow surveys following the recommendations listed by Pomeroy and Gray (1995). 10-point snow surveys under the pine canopy were used in Waskesiu, whilst longer (+100 point) surveys were used in open environments.

Interception was determined from the difference between snowfall and sub-canopy snowfall or from the weight of snow on a suspended pine tree weighed with a load cell (Hedstrom and Pomeroy 1998). To provide areal averages of snowfall, sub-canopy snowfall was compared to the accumulations determined by snow survey during cold periods and adjusted to provide an areal indicator of sub-canopy snowfall. The difference between snowfall and a really averaged sub-canopy snowfall was related to the weight of snow on the suspended tree just after new snow events. The ratio of the difference to weight was used to estimate the canopy snow interception load in mm snow water equivalent.

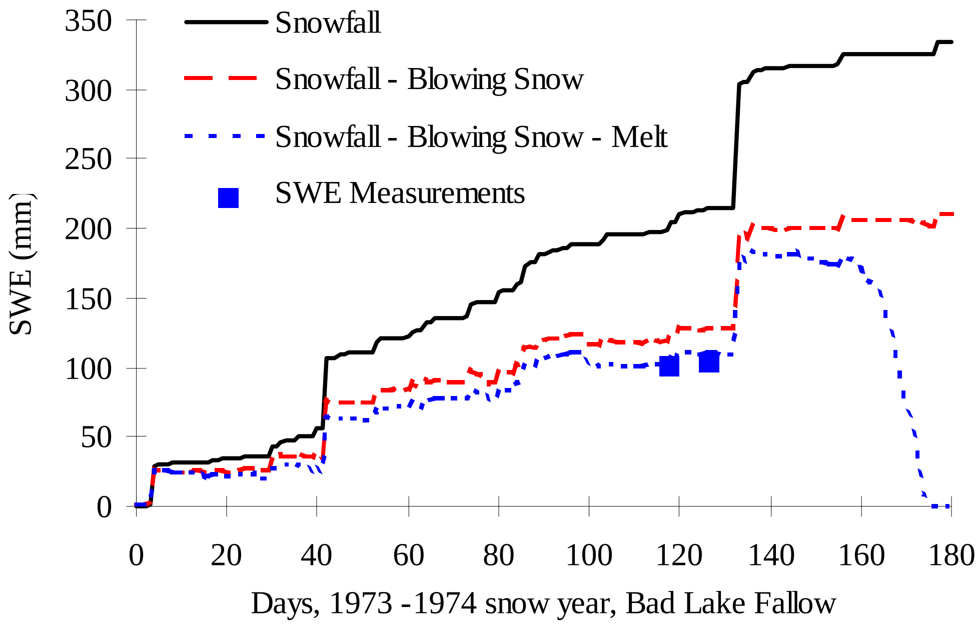
Results

Snow Accumulation and Wind Transport of Snow,

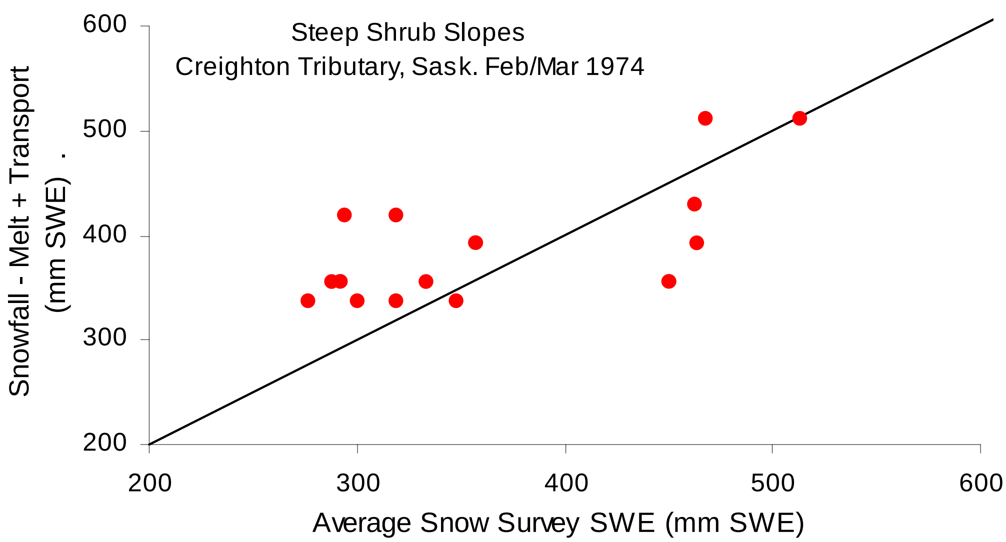
<u>Modelling Blowing Snow</u>

Sublimation fluxes during blowing snow have been estimated to return 10-50% of seasonal snowfall to the atmosphere in North American prairie and arctic environments (Pomeroy and Gray 1995; Pomeroy et al. 1997; Essery et al. 1999). The Prairie Blowing Snow Model (PBSM) (Pomeroy 1989; Pomeroy et al. 1993) has been redeveloped for compatibility with GCMs, incorporating features for independent determination of the transport threshold for drifting (Li and Pomeroy 1997a), upscaling blowing snow fluxes using probability theory (Li and Pomeroy 1997b), improved vegetation parameterizations, simplified calculations for variable fetches and landscape-based snow mass balances that include snowmelt (Pomeroy et al. 1997b; Pomeroy and Li 1997). PBSM simulations for fallow and gully land surfaces, along with the results of extensive landscape-based snow surveys for the Bad Lake basin in 1974 are shown in Figure 1.

The relative agreement in both fallow-field (transport-out and sublimation dominated) and gully (transport-in dominated) landscapes suggests that both the small-scale and large-scale snow mass balances can be simulated by a blowing snow model of this type (transport out of fallow must equal transport into gully and accumulation at either site must also balance any sublimation present). The seasonal sublimation and transport losses over the fallow field were 24% and 15% of annual snowfall presuming a fetch of 1-km, had the fetch been 3-km the sublimation loss would have been 1.5 times this value, i.e., 36%. In the gully, the increases in snow accumulation due to blowing snow transport ranged from 50% to 100% of cumulative snowfall.



(a) Open fallow field, Bad Lake, SK 1973-1974



(b) Steep shrub-covered slopes, Creighton Watershed, SK 1974

Figure 1 **Snow accumulation modelled for: (a) open field of fallow at Bad Lake, SK, 1973-1974 and (b) steep shrub-covered slopes at Creighton Watershed, Sask., 1974, using PBSM along with snow accumulation measured from extensive snow surveys.**

The sublimation fluxes are calculated by PBSM as part of blowing snow two-phase particle transport models with provision for phase change based upon a particle-scale energy balance. Further tests, in addition to the field-scale snow redistribution patterns and snow mass evaluations (above) have been undertaken to authenticate the sublimation routine. Direct evidence has been obtained that large latent heat fluxes (40-60 W m^{-2}) that result in sublimation rates of 0.05-0.075 mm snow water equivalent $hour^{-1}$, are associated with mid-winter, high-latitude blowing snow events (Figure 2). For events with wind speeds above the threshold level for snow transport, these fluxes are in the range of those predicted by the Prairie Blowing Snow Model. The fluxes are well in excess of those found during spring snowmelt and which can be predicted by standard bulk aerodynamic transfer equations.

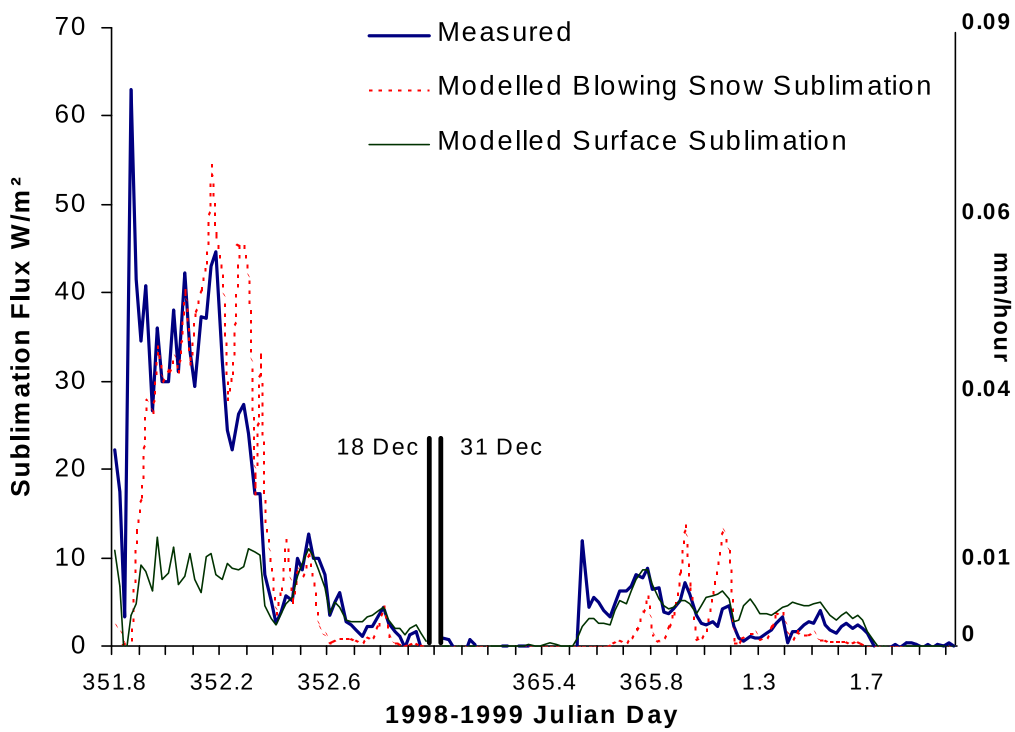


Figure 2 **Measured sublimation flux, modelled blowing snow sublimation (PBSM) and modelled surface sublimation (bulk transfer) measured at over a level Prairie surface.** (after Pomeroy and Essery 1999)

PBSM was simplified, spatially-distributed and driven using a terrain windflow model MS3DJH/3R over the complex arctic terrain of Trail Valley Creek on an hourly time step (Essery et al. 1998). Topography was permitted to vary according to measurements contained in a geographic information system (digital elevation model). Probability of blowing snow occurrence algorithms in the simplified PBSM were sensitive to vegetation type and burial of vegetation by snow, and used to index the effect of variable vegetation roughness on blowing snow. The simulation was run for Trail Valley Creek over the winter of 1996-97 on an hourly time step with a spatial resolution of 80 m. Mean SWE and CV of SWE are within a standard deviation of measured values at the end of the winter season. A simulation with suppressed sublimation provided much greater predicted snow accumulation than that observed. Several new open terrain landscape classes were identified based on wind flow regimes: windswept, windward, divergent, neutral and convergent. The mean SWE and CV of SWE are notably different amongst these regimes.

Figure 3 shows frequency distributions for end of season SWE in various terrain types. The differences are the result of wind redistribution. Normal distributions of SWE, fitted to the frequency histograms predicted by the complex terrain blowing snow model (Figure 3) can be used to initialize models of snowcover depletion. The distributions are also the first physically modelled estimates of the variability of SWE over complex tundra terrain and are critical to "scaling up" snowcover estimates for the Mackenzie basin.

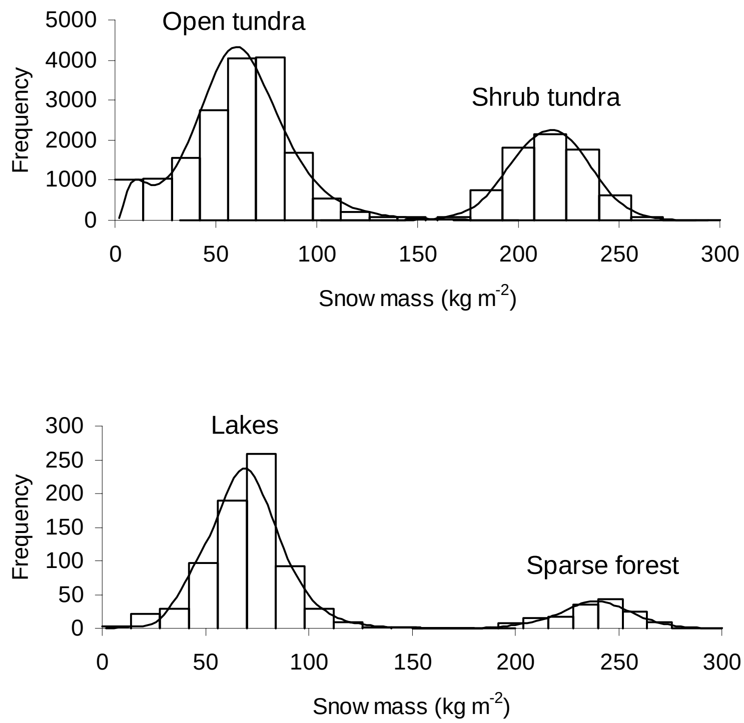


Figure 3 **Histograms showing modelled snow water equivalent (expressed as mass) frequency compared to summed lognormal distributions for terrain types of varied vegetation and topographic placement in Trail Valley Creek, spring 1997.**

Snow Interception

Interception by forest canopies can store up to 60% of cumulative snowfall by midwinter in cold boreal forests, which results in a 30%-40% annual loss of snowcover over the winter in many coniferous forest environments (Pomeroy and Gray 1995). Following interception, most snow remains in the canopy where it may be exposed to a relatively warm and dry atmosphere. As a result, relatively high rates of sublimation occur from intercepted snow, and more snow sublimates than eventually unloads to the surface (Pomeroy and Schmidt 1993; Pomeroy and Gray 1995; Harding and Pomeroy 1996). In order to calculate the amount of sublimation from intercepted snow, it is important to know the amount of snow collected in the canopy so that appropriate exposure times can be determined. For example, given a constant sublimation rate, an underestimation of interception will result in a shorter exposure time for sublimation and a decrease in seasonal sublimation.

Field data on snow interception collected for a pine forest in Beartrap Creek, Saskatchewan is shown in Figure 4. Interception measured by Nipher snow gauges is determined as the difference between weekly above- and below-canopy snowfall for periods when there were no major releases or unloading of intercepted snow. Interception measured via a weighed tree is determined by the increase in weight of snow on a suspended tree, expressed as an areal average snow water equivalent, SWE. It is seen that as much as 9 mm SWE can be intercepted in this canopy and that interception increases with snowfall up to snowfall amounts of about 16 mm SWE, above which interception does not increase further.

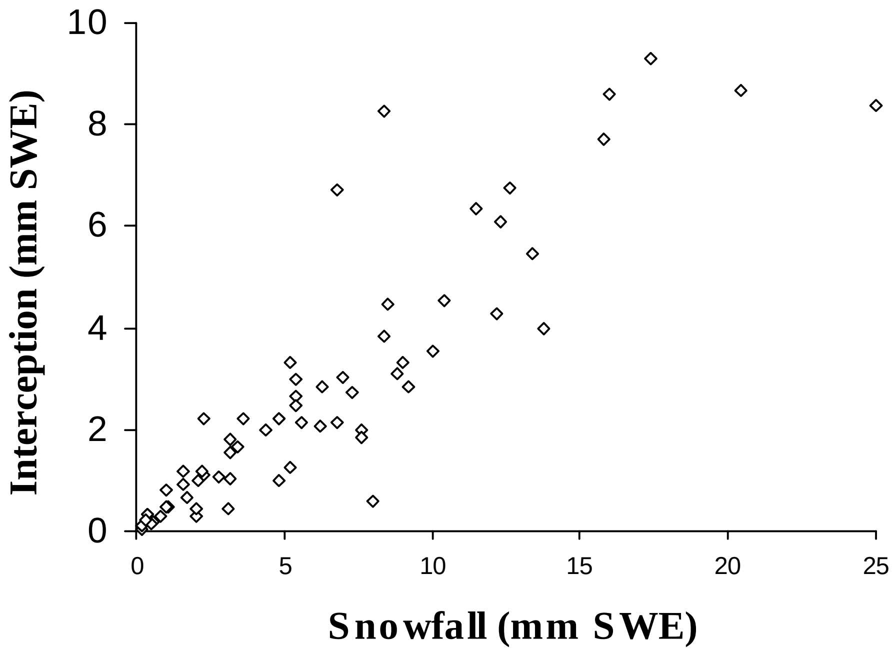


Figure 4 Snow interception of a jack pine canopy in Beartrap Creek, Waskesiu, SK measured using the difference in above and below canopy snowfall, converted to areal SWE using snow survey measurements. (after Hedstrom and Pomeroy 1998)

To model interception in large-scale simulations Pomeroy et al. (1999) recommend a simplification of the routine developed by Hedstrom and Pomeroy (1998). The routine is consistent with field observations that snow interception efficiency decreases with snow canopy load and snowfall amount and increases with canopy density (Pomeroy and Gray 1995; Hedstrom and Pomeroy 1998). Modelling this behavior provides interception, I (kg/m^2), as a function of a dimensionless snow unloading coefficient, c, the difference of the maximum snow load, I^*, and initial snow load, I_o (kg/m^2), an exponential function of snowfall, P (kg/m^2 for a unit time), and the canopy density, C_c (proportional coverage) as (Hedstrom and Pomeroy 1998):

$$I = c(I^* - I_o)(1 - e^{-\frac{C_c P}{I^*}}) \qquad (1)$$

This expression is independent of time step except for the unloading coefficient, c. Empirical evidence suggests a value of c = 0.7 is appropriate for hourly time steps. I^* is found as a function of LAI

(m^2/m^2), a tree species coefficient, S_p (kg/m^2) and fresh snow density, ρ_s (kg/m^3) following Hedstrom and Pomeroy (1998):

$$I^* = S_p LAI\,(0.27 + \frac{46}{\rho_s}) \qquad (2)$$

Schmidt and Gluns (1991) present field measurements that suggest values for S_p of 6.6 for pine and 5.9 kg/m^2 for spruce. The values for coefficients and constants in Eq. 2 are empirically derived and hence based on the units expressed above.

The snow interception routine specifies increasing interception efficiency with LAI and decreasing efficiency with increasing snowfall and initial interception. A comparison of modelled and measured snow interception is shown in (Figure 5).

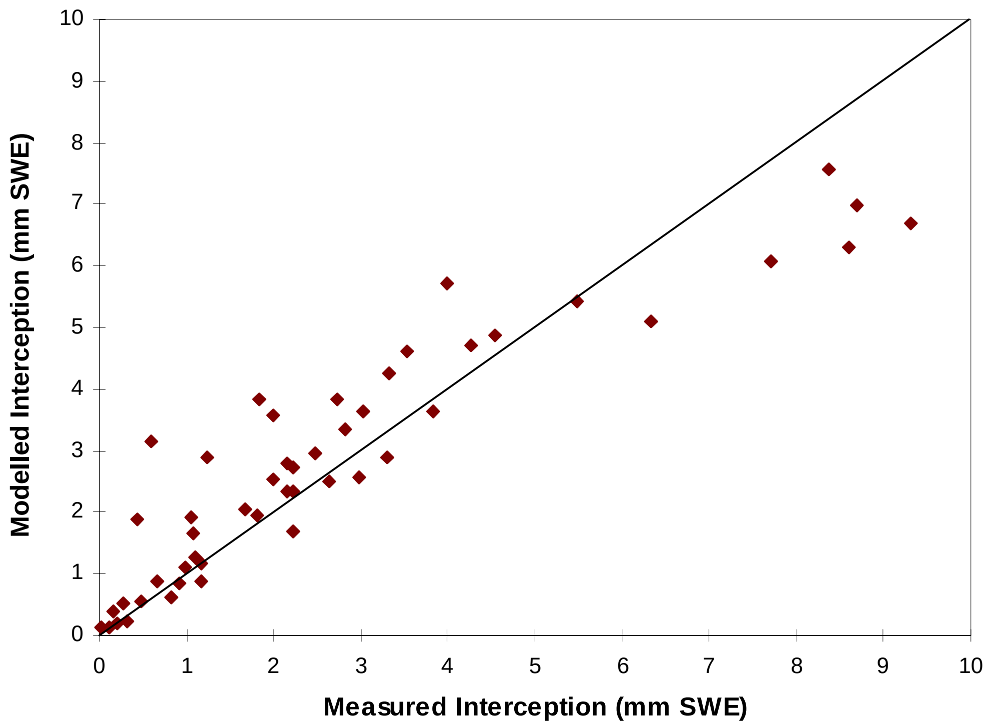


Figure 5 Measured and modelled snow interception. Comparison of the measurements from above and below canopy snowfall gauges, scaled-up using snow surveys and the results of the interception model results for Beartrap Creek, 1993-96.

<u>Sublimation of Intercepted Snow</u>

Physically-based equations describing snow interception (Hedstrom and Pomeroy 1998) and sublimation processes (Pomeroy et al. 1998) were applied to canopy intercepted snow using a fractal scaling technique (Pomeroy and Schmidt 1993) to provide a snow-covered forest boundary condition for a one-dimensional land surface scheme, CLASS (Verseghy et al. 1993). Substantial modifications of CLASS's treatment of turbulent transfer and within-canopy ambient humidity were required to accommodate this nested control volume approach (Figure 6).

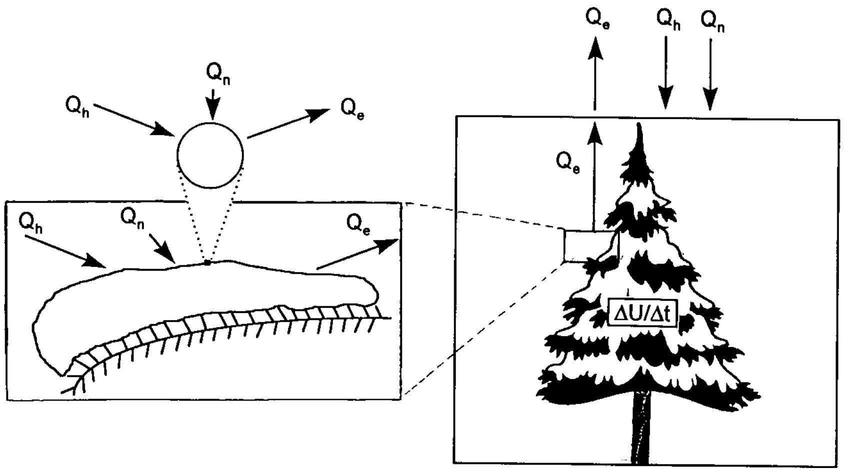


Figure 6 **Nested control volumes for calculation of coupled mass and energy exchange between the atmosphere and intercepted snow. Q_n, Q_h and Q_e are net radiation, sensible heat and latent heat fluxes to particle, branch and tree and ΔU/Δt is the time rate of change of internal energy.** (after Parviainen and Pomeroy, in press)

Tests in late winter in a southern boreal forest against measured sublimation found that the coupled model provides good approximations of sublimation losses on half-hourly and event basis (Figure 7). Cumulative errors in estimating canopy temperature, humidity, and intercepted snow load over 8 days of simulation were -0.7°C, -4.2% of the average observed vapor pressure, and 0.04 kg/m^2, respectively.

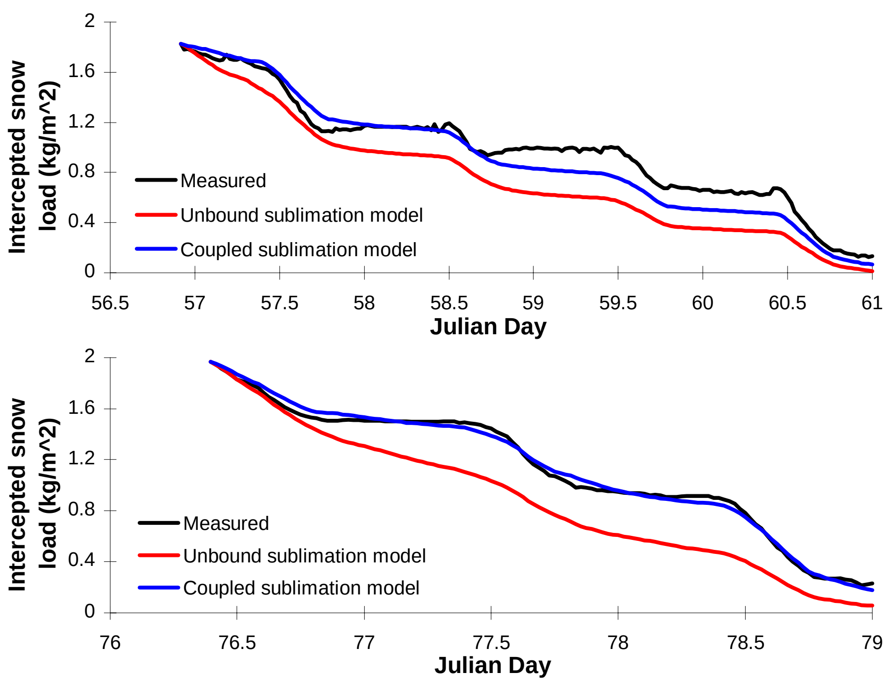


Figure 7 **Measured intercepted snow load (from suspended pine tree), and unbounded sublimation model and coupled land surface scheme sublimation model calculated intercepted load. Unbounded sublimation model driven using snowfall and within canopy measured wind speed, radiation, temperature and humidity. Coupled model driven with snowfall and above canopy reference wind speed, radiation, temperature and humidity.** (Pomeroy et al. 1998)

Testing of the model energy balance against eddy correlation measurements yielded reasonable estimates of latent and sensible heat fluxes during an overnight event, but poorer estimates during periods of large snow loads and sunlight. Further work to incorporate a radiation scaling correction, continuing improvements to heat storage terms, and within-canopy turbulent transfer are expected to improve coupled model performance. These improvements are being presently pursued.

Ablation of Seasonal Snowcover

Open Environments

1. Areal Albedo

The disintegration of shallow snowcover into an areal mosaic of patches of ground and snow during ablation greatly affects the energetics at the land surface and the contributing areas of evaporation, infiltration and runoff. With respect to surface energetics, the major changes are brought about by the decrease in areal albedo that accompanies snowcover ablation. Short-wave energy fluxes are influenced by the snow albedo.

It is often difficult to interpret field measurements of areal albedo, α, made with hemispherical radiometers because of the decrease in snow-covered area during melt. Field measurements of albedo can be corrected using measured fractions of snow-covered area, F_s, from aerial photographs to provide estimates of the variation of α_{sn} during melt. For patchy snowcover, the snow albedo, α_{sn}, is obtained from areal albedo by weighting the albedos of snow and ground according to a segregation based on the fraction of the unit area covered by each surface. This technique is commonly employed by recent land surface schemes to calculate areal albedo during melt, and is:

$$\alpha = \alpha_{sn} F_s + \alpha_g (1 - F_s) \tag{3}$$

where α_g, = the albedo of snow-free ground.

Shook (1993) demonstrated that the association between α and F_s during snowmelt on the Canadian prairies is approximately linear over a large range in F_s (0.9-0.1), as shown in Fig. 8. A linear association between α and F_s suggests that α_{sn} does not vary greatly during ablation and that the major factor controlling the decay in α is the fraction of bare ground. The upper limit of α is established primarily by α_{sn}, closely following the start of depletion in snow-covered area (SCA); the lower limit is set by α_g.

Fitting a linear regression to the data for Smith Tributary, Bad Lake, SK in Figure 8 gives an intercept of 0.096 and a slope of 0.75 with a correlation coefficient, r = 0.99. The corresponding values for the albedos of ground and snow by Eq. 3 are therefore α_g = 0.096 and α_{sn} = 0.85. A similar analyses applied to the measurements for the Creighton sub-basin of Bad Lake gave the corresponding statistics: intercept = 0.095, slope = 0.42, r = 0.95, α_g = 0.095 and α_{sn} = 0.52. The lower albedo for snow on the Creighton Watershed is due to a shallow, dirty snowcover and vegetation protruding through the snow surface. The large differences in snow albedo between snowcover of similar grain size and wetness suggest that much of the variation in albedo with grain size may be overwhelmed by local factors such as dust, pollution and protruding vegetation. These results also suggest that the changes in the reflectance of snow during ablation are small and have little effect on α. The data call into question the presumption in albedo calculations that melting snow is "clean" and covers vegetation during melt.

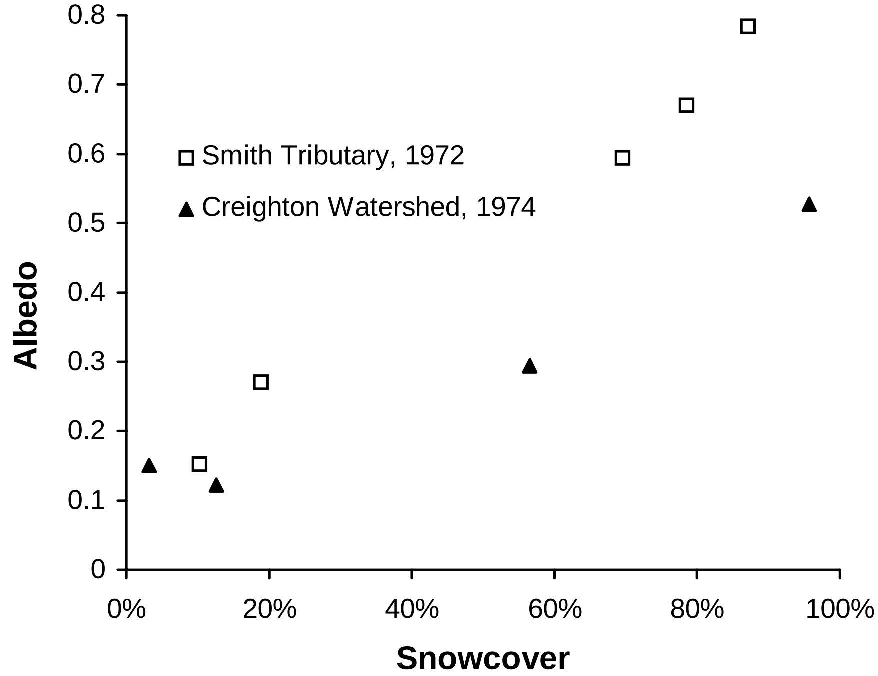


Figure 8 **Relation between areal albedo, α_A, and snow-covered area, SCA, monitored on two small watersheds at Bad Lake, SK.**

2. Snowcover Depletion

Shook (1993, 1995) and Shook et al. (1993, 1994) demonstrated that in environments where the melt flux is relatively uniform over an area, the fractal character of the spatial distribution of the water equivalent controls the geometry's of the soil and snow patches of an ablating snowcover. He found that the spatial frequency distribution of the snow water equivalent (SWE), of a natural snowcover often can be approximated by the lognormal probability density function, which can be described by the linear form:

$$SWE = \overline{SWE}\,(1 + KCV) \tag{4}$$

where: SWE = snow water equivalent having an exceedance probability equal to that of the frequency factor, K (Chow 1954),

$\overline{\text{SWE}}$ = mean snow water equivalent, and
CV = coefficient of variation

Figure 9 plots point measurements of SWE and fitted lognormal distributions for three landscapes in a prairie environment: (a) a relatively flat field in wheat stubble, (b) an undulating field of fallow and (c) a relatively flat low area with scattered brush. The respective coefficients of determination, r^2, between the measured and fitted SWE-values for the landscapes are 0.92, 0.98 and 0.99, respectively.

Shook (1995) also noted that the snowcover-depletion curve (SDC) is produced by applying the snow melt rate evenly over the SWE frequency distribution. Under conditions where the melt flux over an area is reasonably uniform, fair representation of the depletion of snow-covered area (SCA) during ablation can be obtained by melting the frequency distribution of SWE. The results of this procedure are demonstrated in Figure 10, which plots the decrease in SCA over time for a constant melt rate, assuming initial conditions of a lognormal distribution of SWE with various CV values. The results show that the smaller the CV, the more rapid the depletion in SCA. This trend is the result of the increased peakedness of the frequency distribution of SWE with decreasing CV. The smaller the CV, the larger the number of SWE-values grouped near to the mean.

The snow-covered area depletion algorithm developed by Shook (1995) was tested using data collected on the Smith Tributary at Bad Lake in 1972, with melt rates driven by a temperature-index model. The results are shown in Figure 11 and demonstrate reasonable correspondence between modelled and measured areal depletion of snowcover (Shook 1995).

Shook and Gray (1997) developed a method of synthesizing a snowfield from point field measurements based on the fractal structure of the snow water equivalent. The synthesis involves two steps: (a) using the method of the fractal sum of pulses to generate the spatial distribution of snow water equivalent in an array and (b) adjusting the synthetic snow field to take on the statistical properties of the natural snowcover, as described by the two-parameter lognormal probability density function fitted to field data (e.g., Eq. 4).

Figure 12 shows close agreement of the perimeter-area relationship of ablated snow patches for synthetic and natural snow fields on Smith Tributary, a 1.9 km^2 in the prairie region of Saskatchewan, Canada (Lat. 51^o 19'N, Long. 108^o 25'W), when the watershed is about 62% snow-covered. The fractal dimensions of the synthetic and snowcover were 1.36 and 1.35, respectively.

The methodologies for deriving areal snowcover depletion curves described above require information on the mean depth and standard deviation or coefficient of variation of the snow water equivalent (see Eq. 4). On the presumption that reasonable estimates of the mean depth would be available from snow accumulation models applied to point snowfall measurements, considerable effort during MAGS was been placed on the measurement and tabulation of the coefficient of variation of snow water equivalent, C_{SWE}, for different landscapes.

A summary of average values for C_{SWE} for various landscapes in Prairie, Arctic and Boreal Forest environments is given in Table 1. It is important to note that these values assume random sampling.

Shook and Gray (1996) found that, in shallow snowcover where there is strong autocorrelation between snow depths at adjacent measurement points and weak covariance between snow depth and density, the standard deviation increasing with sampling distance up to some maximum, the “cutoff” or “correlation” length (see Figures 13a and 13b). Measurements at greater sampling intervals follow random behavior in which the standard deviation and coefficient of variation tend to reasonably constant values.

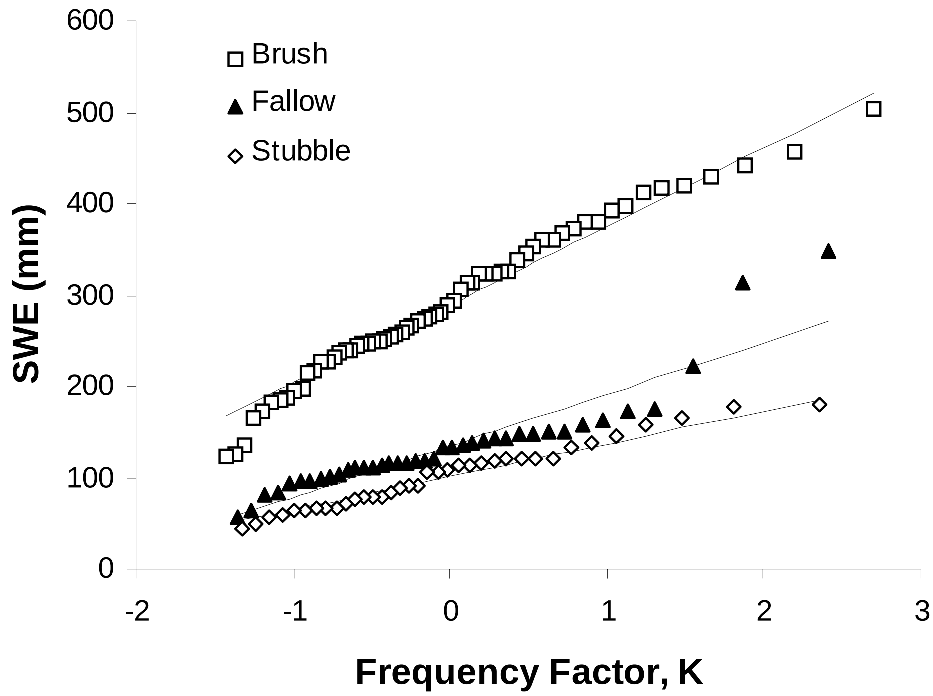


Figure 9 **Measurements of SWE, fitted lognormal distributions and associated frequency factor K, for three landscapes. The r^2 values for the measured and fitted SWE values are 0.92 for stubble, 0.98 for fallow and 0.99 for brush.**

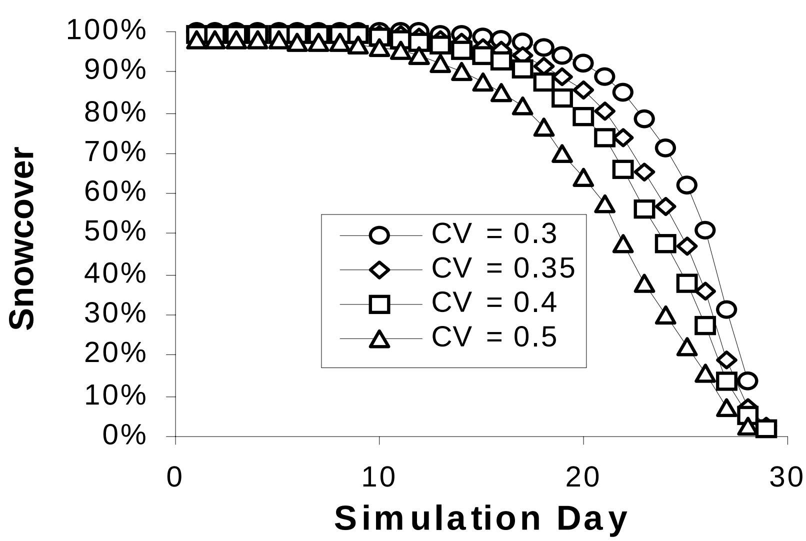


Figure 10 **Snow-covered area depletion curves (SDC) modelled by applying a constant, uniform melt rate to snowcover whose SWE have a log normal distribution. SDC is plotted against time from an initial SWE of 130 mm for various coefficients of variation of SWE.**

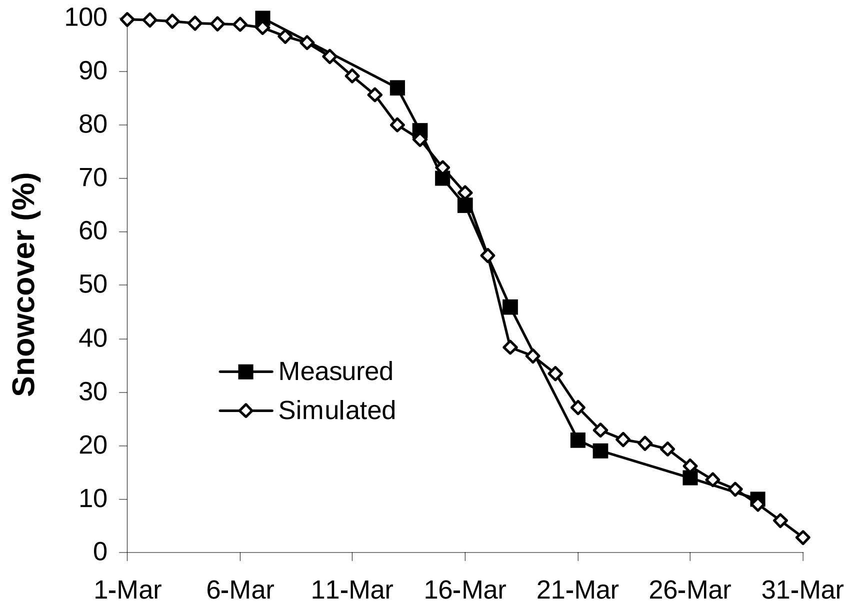


Figure 11 **Modelled and measured snowcover depletion curves during snow ablation on the Smith Tributary in 1972.**

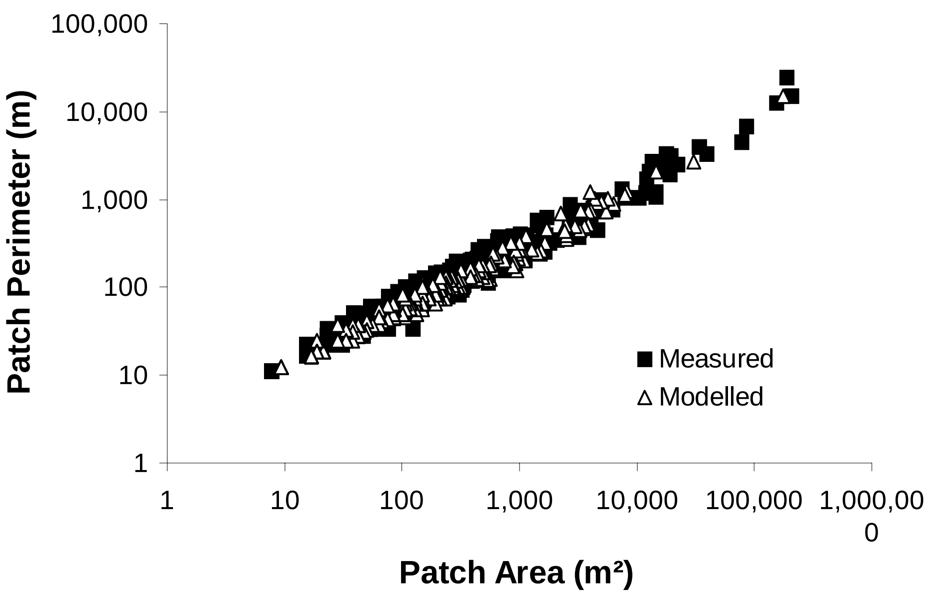


Figure 12 **Perimeter – area relationships of snow patches for synthetic and natural snowcover.**

Table 1 **Representative values for the coefficient of variation of water equivalent of snowcover, CV_{SWE}, on various landscapes in Prairie, Arctic and Boreal forest environments in late winter.**

Region	**Landuse and Vegetation**	**Landform**	**CV_{swe}**
		Flat plains; Slightly to moderately rolling topography with gentle slopes	0.47
	Fallow	Bottom (bed) of wide waterways; Large sloughs and depressions	0.30
		Crests of hills, knolls, and ridges	0.58
PRAIRIES	Stubble	All Landforms	0.33
		Flat Plains; Bottom (bed) of wide waterways; Large sloughs and depressions; Slightly to moderately rolling topography with gentle slopes	0.41
	Pasture	Crests of hills, knolls, and ridges	0.51
		Lee of abrupt, sharp slopes	0.57
	Scattered Brush	Bottom (bed) of waterways, e.g., gullies; Sloughs and depressions	0.42
		Lee of abrupt sharp, slopes	0.52
	Treed Farm Yards		0.50
		Flat plains, upland plateaus, slight to moderately rolling topography	0.31
	Tundra	Valley bottoms	0.28
		Valley sides (drifts where slopes greater than 9^o)	0.34
		Flat plains, slight to moderately rolling topography	0.22
ARCTIC	Shrub Tundra	Valley bottoms	0.16
		Valley sides (drifts where slopes greater than 9^o)	0.18
	Sparse	Exposed hillside and forest edge	0.21
	Forest-tundra	Sheltered	0.11
	Black Spruce		0.14
	Mature Jack Pine		0.10
BOREAL	Mixed Aspen-White Spruce	All Landforms	0.05
FOREST	Young Jack Pine		0.14
	Recent Clear-cut		0.07
	Recent Burn		0.04

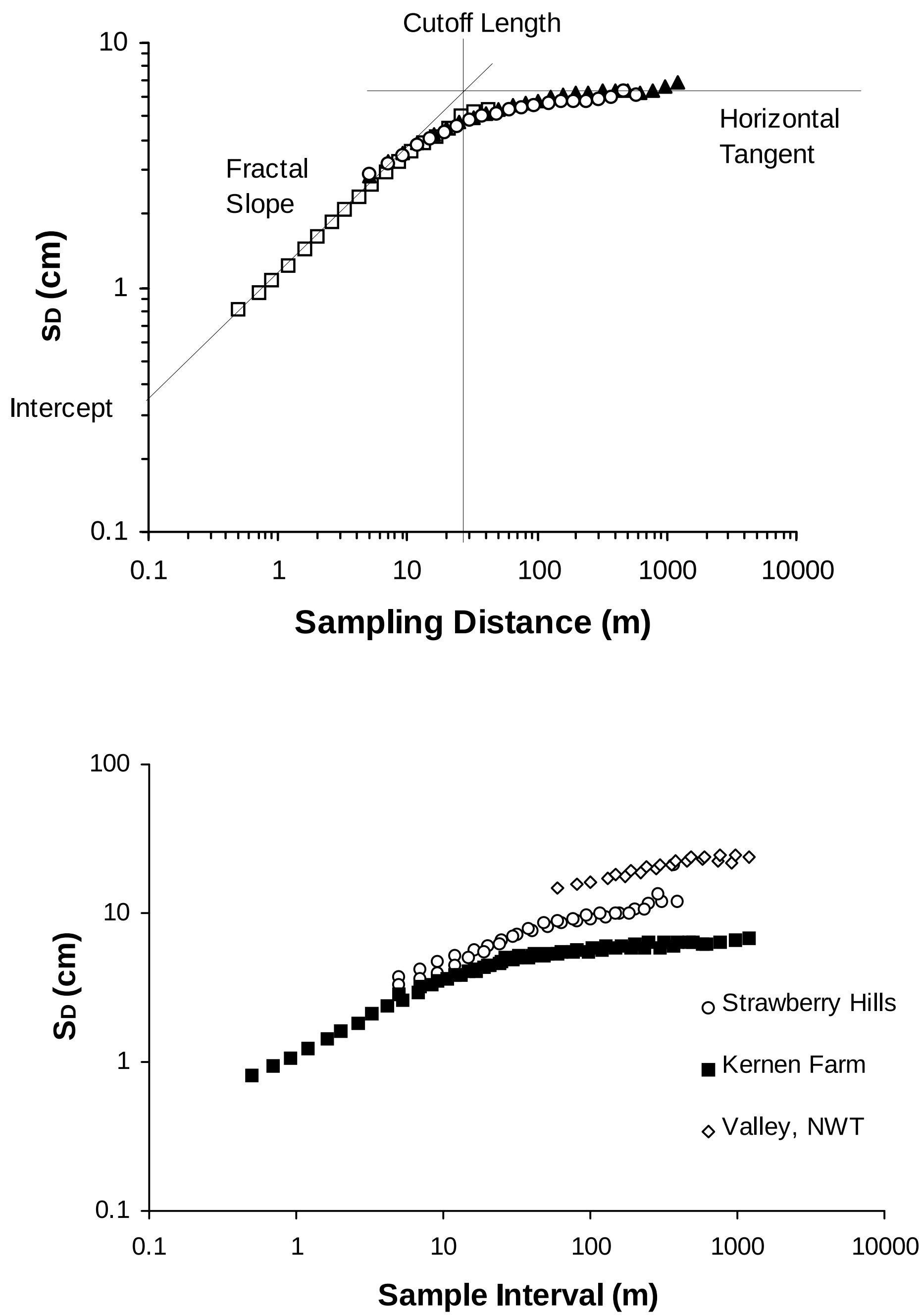


Figure 13 **Variation in standard deviation of snow depth, s_D and sampling distance: (a) The transition in snow depth from fractal to random structure as defined by the he intersection of the fractal slope and horizontal tangent to the points; (b) data from Kernen Farm and Strawberry Hills near Saskatoon, SK and Trail Valley Creek Watershed near Inuvik, NWT. The approximate "cutoff" lengths for the three areas are 30 m, 80 m and 500 m, respectively.**

Boreal Forest Environment

The influence of forest canopy cover and variable melt energetics on depletion of snowcover was investigated in the boreal forest of central Saskatchewan following earlier work by Shook (1995) and Shook and Gray (1996) in open environments. The results can be distinguished between variability within the forest stand and that between forest stands. Within stands, Faria (1998) found the frequency distribution of SWE (snow water equivalent) under boreal canopies was found to fit a log-normal distribution, with the most dense stand displaying the most variable SWE prior to melt. Higher variability in SWE results in earlier exposure of ground under spatially uniform melt conditions (Shook and Gray 1996), but within stands, snowmelt energy below the canopies was found to be spatially heterogeneous and inversely correlated to SWE (Figure 14). The variability of melt energy within a stand decreased with overall stand density.

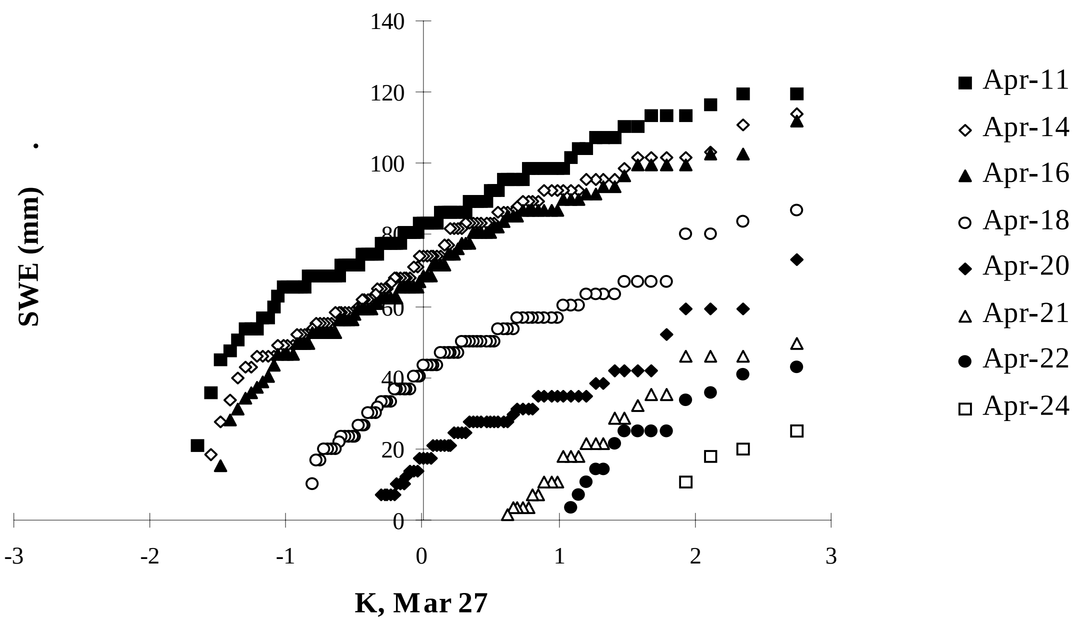


Figure 14 Sequential distributions of snow water equivalent (SWE) during melt in a Pine Stand. K is the frequency factor for the lognormal distribution of SWE), the K for SWE=0 reflects snow covered area. Note that melt is greater for smaller SWE. (after Faria et al. in press)

Within-stand covariance between the spatial distributions of snow water equivalent and melt energy promoted an earlier depletion of snowcover than if melt energy were uniform. This covariance was largest for the most heterogeneous stands (usually medium density). Stand scale variability in mean SWE and mean melt energy resulted in more rapid SCA depletion in those stands having the lower leaf area. Because of the heterogeneity in the spatial distributions of SWE and melt energy in forest environments, it is necessary that these variations be included in calculations of snow covered area (SCA) depletion (Faria et al. in press). Figure 15 shows an example calculation where initial SWE and mean melt energy at the stand scale are used to drive SCA depletion calculations which rely on the initial sub-stand distribution of SWE and the covariance between SWE and melt. Comparisons of the measured depletion with simulated depletion showed improved fit for simulations that included covariance over those that neglect this feature (Faria et al. in press).

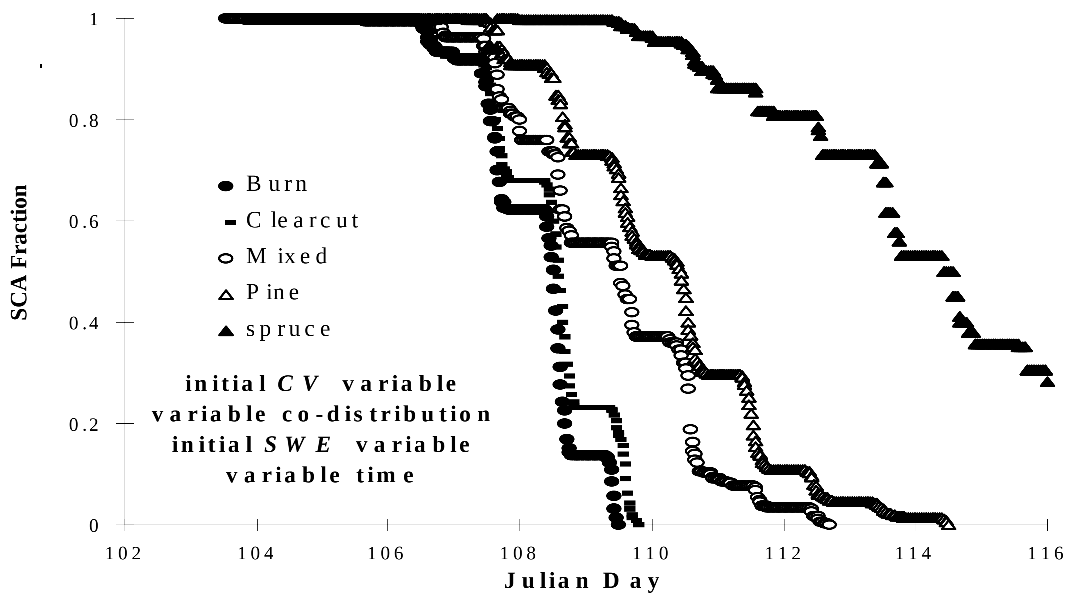


Figure 15 Simulated snowcover depletion curves using measured mean melt rate at each site to calculate change in snow covered area with time as a function of the distribution of SWE and the covariance between SWE and melt rate. (after Faria et al. in press)

Coupled Heat and Mass Transfer in Snow and Underlying Frozen Ground

Numerical Model

The studies of coupled heat and mass transfer conducted within MAGS were directed to the problem of infiltration into frozen ground and combined a theoretical analyses and field measurement program. A physically-based numerical model, HAWTS (Heat And Water Transport in Frozen Soils) (see Zhao *et al*. 1997; Zhao and Gray 1997a; Zhao and Gray 1998), was developed and used to calculate moisture movement associated with sensible and latent heat transfers in frozen soils, is described in earlier papers.

The model assumes: (a) the soil matrix may be treated as a homogeneous porous medium; (b) the various soil phases are in thermal equilibrium; (c) volumetric changes within the soil matrix can be neglected; and (d) phase changes involving the vapor phase are small during snowmelt. The problem is formulated as a case of one-dimensional, transient, simultaneous mass and heat flow in an unsaturated frozen soil with phase change. The energy equation includes the contributions of conduction, convection and latent heat. Although the convective heat transfer rate is small, the cumulative contribution over time by convection to the soil heat balance should not be neglected (Zhao et al. 1997).

The physical data used in the study of infiltration included field measurements of total soil moisture and soil moisture changes, solar radiation, air, snow and soil temperatures, and snow depth and density. Profiles of total soil moisture (ice + water) changes were calculated from measurements made with a twin-probe gamma system and used to estimate snowmelt infiltration and to establish initial soil saturation and initial soil surface saturation. Complete details of the procedures followed in the soil moisture measurement program are described by Granger *et al*. (1984), and Gray and Granger (1986, 1990). Incoming solar radiation was monitored by Kipp and Zonen pyranometers; air, snow, and soil temperatures were measured using Omega type E thermocouples. All radiation and temperature data were recorded at 30-minute intervals. Weekly measurements of snow depth and snow density were obtained using a ruler and ESC-30 snow density gauge.

Model Performance

Figure 16 compares the model performance for two distinct environments and soil types. The change in volumetric moisture content is plotted with depth for the numerical model and compared to field measurements. Field data were collected during snow ablation from Mar 17 to 19 in a prairie environment in a heavy clay soil (unfrozen saturated hydraulic conductivity of 0.11 cm h^{-1} (Miller 1994; Toth 1999); and from Mar 25 to 26 within a boreal forest environment in a sandy loam soil (unfrozen hydraulic conductivity of 2.28 cm h^{-1} (Elliott et al. 1998). Field measured profiles of soil temperature and moisture were used as initial distributions while measurements of air temperature and solar radiation were used to determine when snow melt was occurring. Average soil surface saturation, S_o, during infiltration was taken as the average moisture saturation of the top two centimeters of soil on the second day. Average surface saturation was 0.60 and 0.59 for the clay and sandy loam soils, respectively, and initial surface soil moistures for the top 30 cm were 30.2 (0.49 saturation for the clay soil) and 28.1 per cent by volume (0.46 saturation for the sandy loam soil). The data in Figure 16 show good agreement between simulated and measured profiles. Total simulated and measured infiltration amounts during the measurement periods were estimated as 5.8 mm compared to 6.5 mm for the clay soil and 15.1 mm compared to 15.0 mm of infiltration in the sandy loam soil.

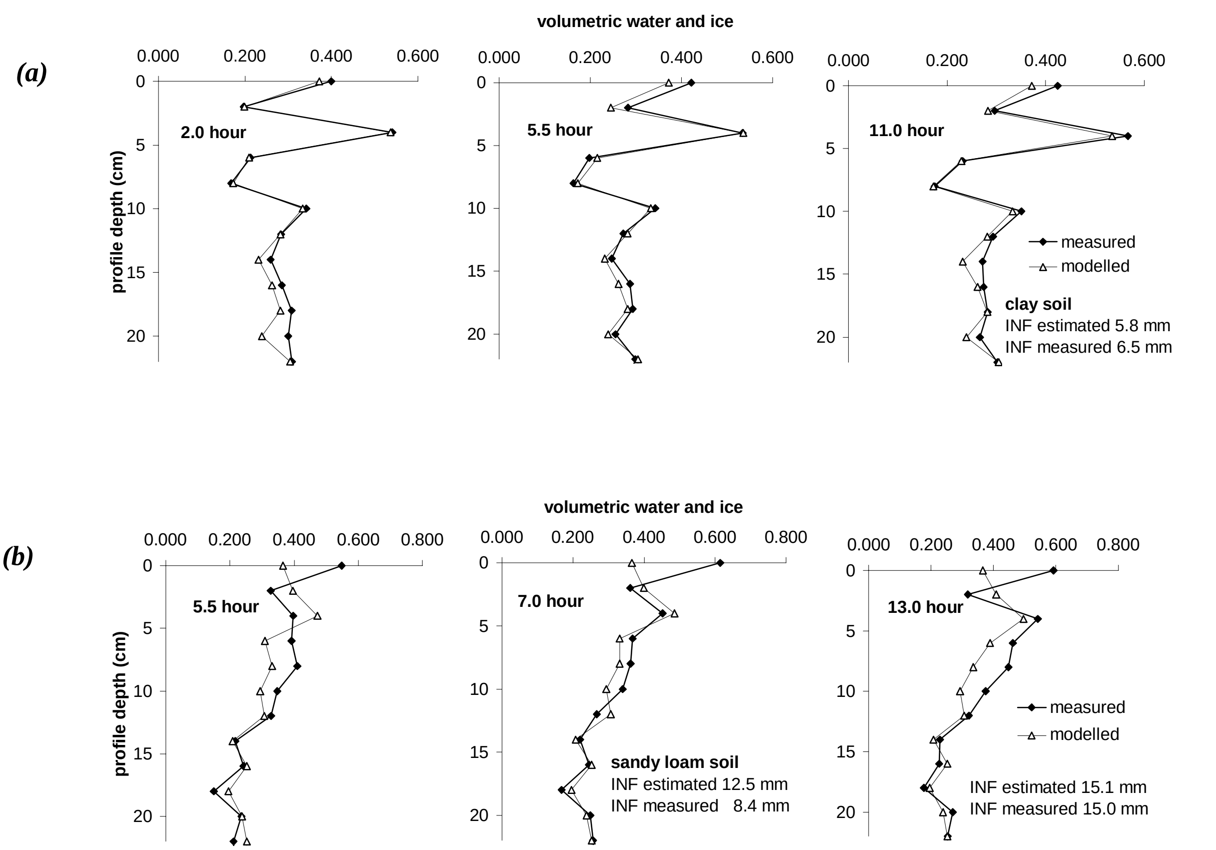


Figure 16 Modelled and measured profiles of soil moisture (water + ice) into a frozen silty clay soil at the Kernen Farm after 2 h, 5.5 h and 11.0 h of snowmelt infiltration simulation, silty clay soil. Simulation initiated at 1200 h March 17/99 and compared to measurements of soil moisture (water + ice) at: 1700 h March 17/99, 1630 h March 18/99 and 1100 h March 19/99; (b) Comparison of modelled and measured profiles of soil moisture (water + ice) into a frozen sandy loam soil at the Prince Albert Model Forest after 5.5 h, 7.0 h and 13.0 h of snowmelt infiltration. Simulation initiated at 1200 h March 25/99, 0900 h March 26/99 and 1800 h March 26/99.

Infiltration Process

Infiltration into frozen ground involves simultaneous coupled heat and mass transfers with phase changes. Field measurements (Kane and Stein 1983) and model simulations (Zhao et. al. 1997) demonstrate that both the infiltration rate and the surface heat transfer rate (conduction) decrease with time following the application of meltwater to the surface (Figure 17). Zhao et. al. (1997) suggested that two regimes, a transient regime and a quasi-steady state regime could describe these variations. The transient regime follows immediately the application of water to the soil surface. During this period the infiltration rate and the heat transfer rate decrease rapidly. The quasi-steady state regime develops when the changes in the infiltration rate and the heat transfer rate with time become relatively small. The duration of the transient period is usually short (a few hours), depending on the hydraulic properties of the soil, surface condition, initial water content and initial temperature of the soil. During transient flow, the energy used to increase the soil temperature is largely supplied by heat conduction at the surface (high heat transfer rate at the surface) (Zhao et. al. 1997). In the quasi-steady state regime, the energy used to increase the soil temperature at depth is supplied by latent heat released by the refreezing of percolating meltwater in the soil layers above (low heat transfer rate at the surface).

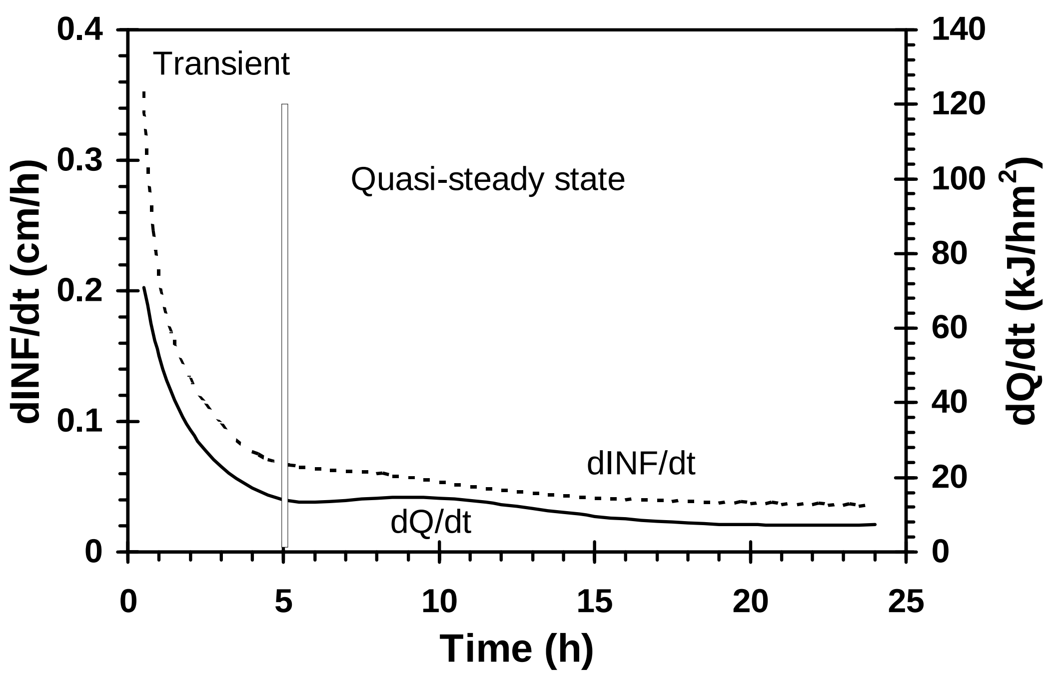


Figure 17 Variations in infiltration rate, dINF/dt, and surface heat flux rate, dQ/dt, with time during snowmelt infiltration into a frozen silty clay soil.

The rapid and dramatic alterations to the soil temperature gradient in frozen soils during snowmelt infiltration caused by the release and downward conduction of latent heat due to freezing of percolating meltwater can only be described properly by a multi-layered soil model having a reasonably-small grid spacing. For those land surface models with only a few soil layers, the ground heat flux during infiltration into frozen soils should not be calculated by the temperature gradient approach. Instead, it is likely that the assumption of a very small value for ground heat (e.g., 0 to 5 W/m^2) or the use of parametric or empirical correlations for estimates of both the ground heat and infiltration will give better results.

Simulating Infiltration into Frozen Soils in Operation Models

Gray et al. (1986) proposed that frozen soils could be separated into three general *"Groups"* in terms of their infiltrability for snowmelt:

Unlimited: soils that are capable of infiltrating most or all available meltwater; this group includes:

- Dry, severely cracked colloidal soils – usually heavy-textured mineral soils under vegetation during the summer months.
- Dry, coarse-textured and gravelly soils – deep, relatively-uniform profiles and sand, sandy loam or other permeable material underlying well-drained peat and organic deposits (e.g., forest soils).
- Sloping, shallow, highly permeable deposits underlain by a relatively impermeable subsurface material (e.g., forest litter and moss on bedrock, tundra – peat on permafrost).

Restricted: soils whose infiltrability is restricted by an impervious surface; this group includes:

- Mineral and other soils frozen in a wet condition (e.g., >70 – 80% pore saturation).
- Soils in which an impeding layer develops at/near the ground surface. For example, a basal ice lens that forms at the soil surface due rain that occurs near freeze-up or the freezing of percolating or infiltrating meltwater.

Limited: soils whose infiltrability is governed primarily by the soil moisture content (water + ice) and soil temperature at the start of snow ablation and the infiltration opportunity time.

Formulations for Limited Soils

Investigations on the problem infiltration into frozen soils focussed on gaining an understanding of the process and the development of an algorithm for estimating infiltration into soils of the Limited Class (Zhao and Gray 1997a and b, 1999; Zhao et al. 1997; Gray et al. 2000). These investigations led to the parametric equation for estimating infiltration, INF (mm), as a function of infiltration opportunity time, t_o (h), and pertinent soil variables:

$$\mathrm{INF} = C S_o^{2.92} \left(1 - S_I\right)^{1.64} \left(\frac{273.15 - T_I}{273.15}\right)^{-0.45} t_o^{0.44}, \tag{5}$$

where: C = coefficient,
S_o = surface saturation - moisture content at the soil surface (mm^3/mm^3),
S_I = average soil saturation (water + ice) of 0-400 mm soil layer at the start of infiltration (mm^3/mm^3). $S_I = \theta_I/\phi$, where θ_I = average volumetric soil moisture (water + ice) at start of infiltration (mm^3/mm^3) and ϕ = soil porosity (mm^3/mm^3), and
T_I = average temperature of 0-400 mm soil layer at the start of infiltration (K)

The coefficient C characterizes the effects on infiltration of differences between model and natural systems. For example, the expression (Eq. 5) assumes quasi-steady flow (Figure 17) when surface saturation is constant, a soil is homogeneous, and the vertical distributions of soil moisture and soil temperature at the start of infiltration are uniform and constant.

Figure 18 compares calculated and measured seasonal infiltration for various (sandy loam, loam, silty clay and clay) fine-textured frozen Prairie soils for frozen sandy soils in a boreal forest at various levels of surface saturation. For the Prairie soils C = 1.85 with a standard deviation of difference between calculated and measured values, s_d = 10.5 mm and for the forest soils C = 0.99 and s_d =12.1 mm with So = 1.0. C can be adjusted for a decrease to surface saturation as: $C_{S_O} = C_1 / S_o^{2.92}$, where C_{S_O} = coefficient for surface saturation, S_o, and C_1 = coefficient for S_o = 1.0.

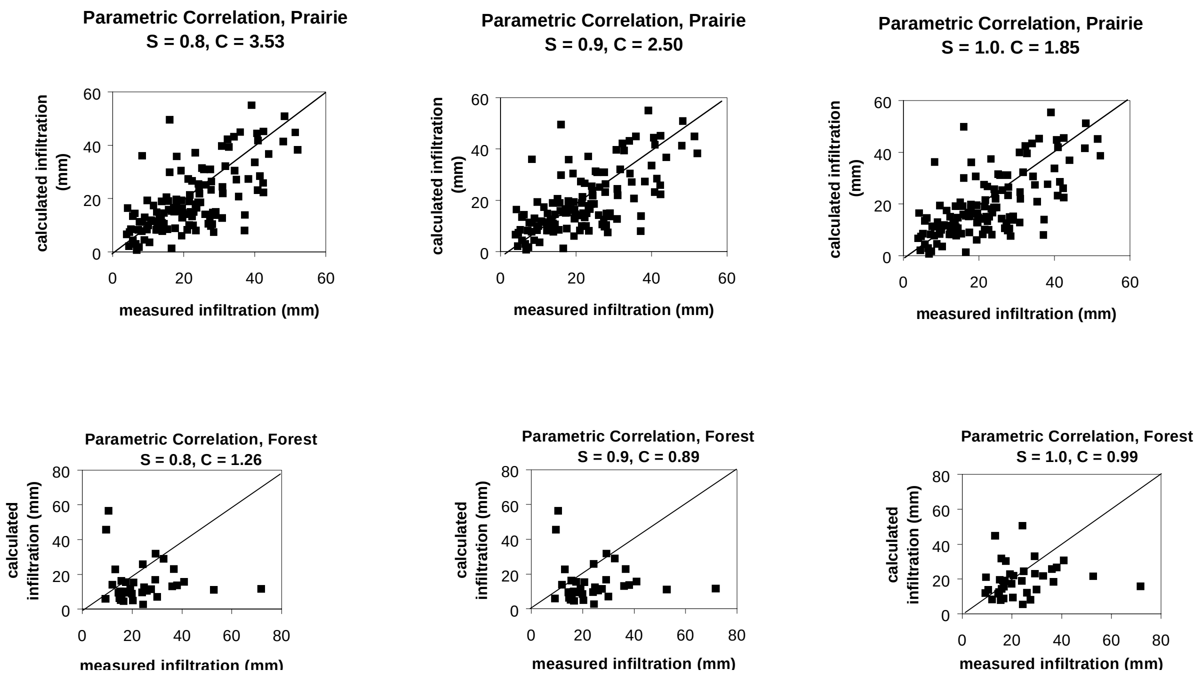


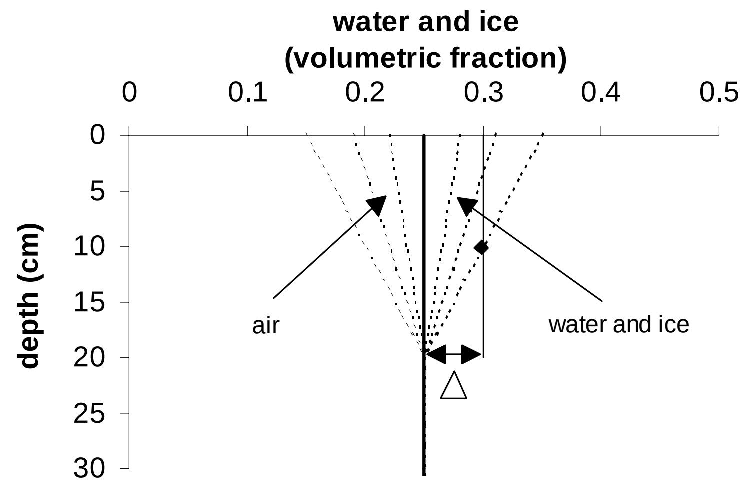
Figure 18 Seasonal infiltration calculated by parametric expression (Eq.5) using various levels of surface saturation, S_0, and "best-fit" values for C versus measured infiltration: (a) Prairies and (b) Boreal forest.

The possibility that C may vary substantially is a serious limitation to the use of Eq. 5 for estimating for estimating frozen soil infiltration, especially if the cause(s) of the variation remain unknown. One of the factors investigated was the vertical distribution of soil moisture (water + ice) in the surface layers of soil at the start of infiltration. Variations in the vertical distribution of moisture cause the average value for S_I (Eq. 5) to change from the corresponding value for a uniform distribution. That is, when the moisture content increases with depth the average S_I is smaller than the corresponding value for a uniform distribution. Conversely, a profile in which the moisture content decreases with depth causes the average S_I to be larger than the corresponding value for a uniform distribution.

In most frozen soils a vertical gradient in soil moisture (water + ice) does not usually result in a significant gradient in soil suction, especially at temperatures lower than about -0.5^{o}C, because the unfrozen (liquid) water content changes slowly with decreasing temperature this value. Thus, the primary effect of a decrease in S_I or θ_I (volumetric content) on infiltration is due to the reduction in ice content, which increases the air-filled porosity, the soil permeability and the infiltration rate. An increase in S_I or θ_I has the opposite effect since an increase in ice content causes the air-filled porosity, the soil permeability and the infiltration rate to decrease.

A series of trials were conducted with the numerical model, HAWTS, (Zhao et al. 1997; Zhao and Gray 1997, 1999) to study the effects of the vertical distribution of soil moisture (water + ice) on cumulative infiltration. These tests assumed a linear variation in soil moisture that extended from the soil surface to various depths (see Figure 19a).

(a)



(b)

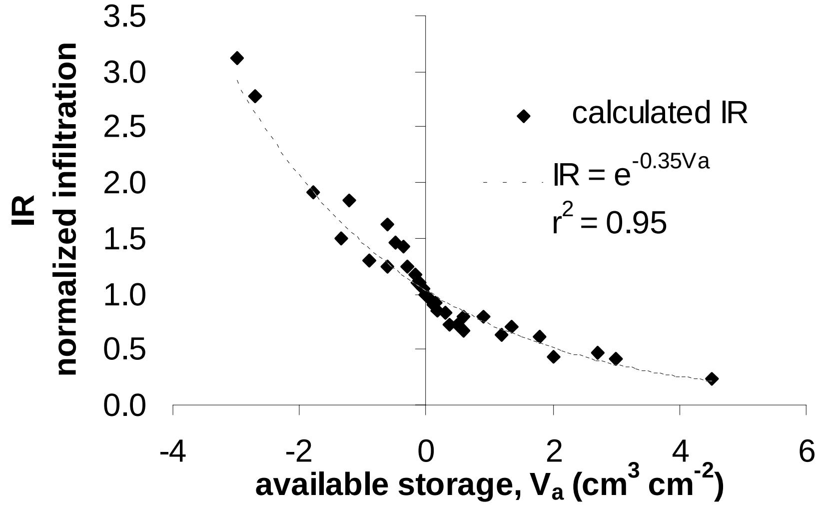


Figure 19 (a) Assumed linear variation in soil moisture extending from the soil surface to various depths: (b) Infiltration ratio, IR, is plotted against the available storage, V_a (i.e. the average difference in volumetric soil moisture (water + ice), Δ, multiplied by the depth of wetting). Negative values of V_a represent a profile in which the moisture content is increasing with depth and positive values of V_a represent a profile in which soil moisture content decreases with increasing depth.

Cumulative infiltration after 12 h was calculated for each distribution. This amount was normalized to the cumulative infiltration after 12 h assuming that the soil moisture was distributed uniformly. The ratio, referred to as the infiltration ratio, IR, is plotted against average difference in volumetric soil moisture (water + ice), Δ, multiplied by the depth of wetting) due to the distribution of soil moisture. Negative values of V_a represent a profile in which the moisture content is increasing with depth and positive values of V_a represent a profile in which soil moisture content decreases with increasing depth.

The data in Figure 19b show that the infiltration ratio IR is sensitive to the vertical distribution of soil moisture. Drier soils near the surface enhance infiltration and produce values of IR > 1. Conversely, wetter soils near the surface suppress infiltration-causing IR < 1. Tests have shown that the curves are independent of the other variables that comprise the parametric equation (Eq. 5). The association between IR and V is described by the expression:

$$IR = e^{-0.35\, V_a} \tag{6}$$

coefficient of determination, $r^2 = 0.95$. The findings confirm the importance of the vertical distribution of soil moisture on infiltration into frozen soils. Large discrepancies among measured infiltration and values estimated by Eq. 5 can be expected if the effect is not accounted for in the calculations.

Application of Infiltration Model

Using the findings above, Gray et al (2000) proposed the routine shown in the flow chart (Figure 20) for implementing the infiltration model in operational systems. The first step involves partitioning a response unit (e.g., catchment, hydrological response unit or other) first into landscape units according to their snow accumulation characteristics. The infiltration model is then applied to each individual landscape unit that is frozen and receives water from melting snow. Further division of the landscape units is recommended depending on the study objective and the sensitivity of the simulation spatial changes of a variable. For example in studies where small scale (local) advection, contributing area and soil moisture recharge are important factors.

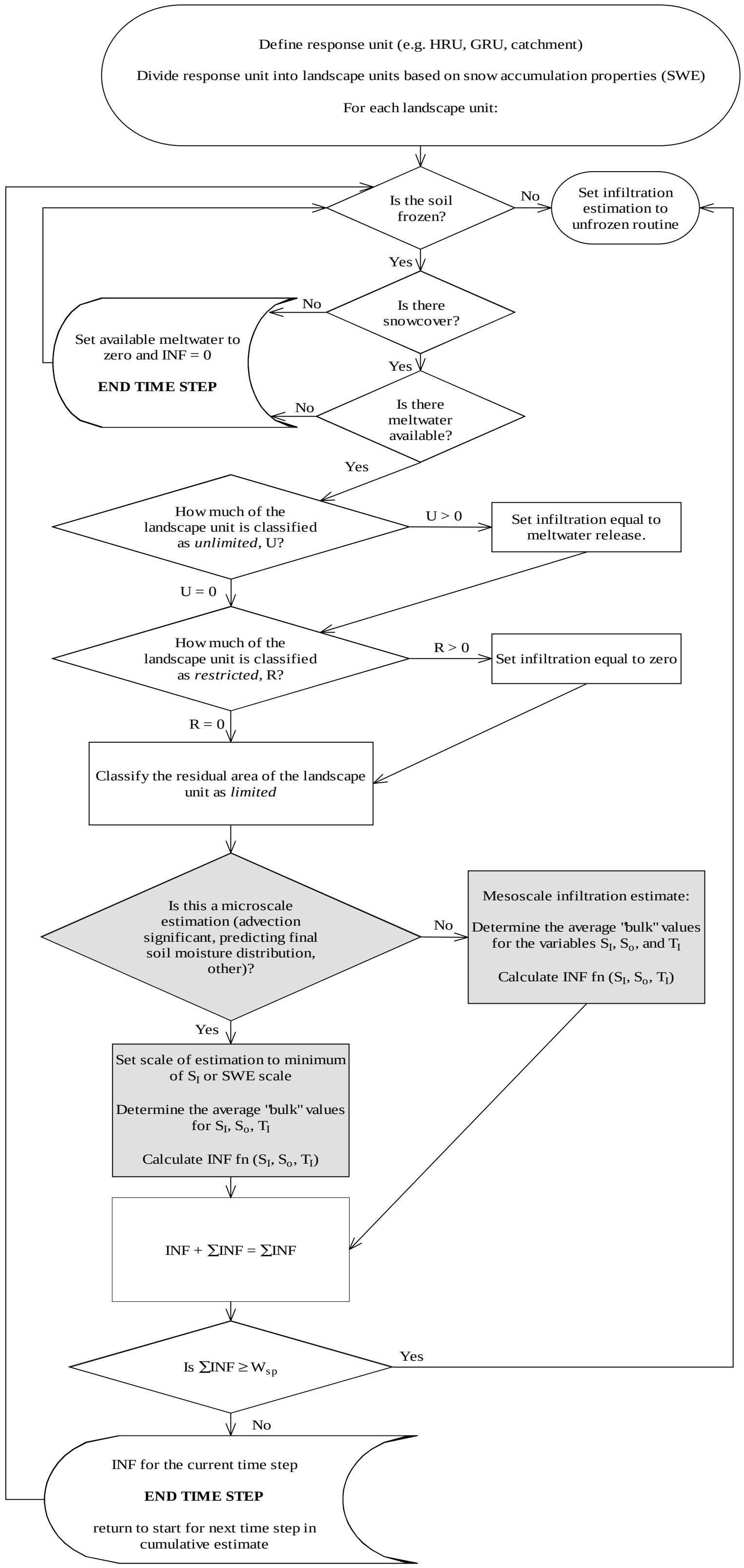
An important point of the convention, which is not discussed above, is restraining the cumulative infiltration calculated by the parametric equation the water storage potential of a frozen soil, W_{sP} (mm):

$$W_{sP} = \phi(1 - S_I)z_f\,. \tag{7}$$

where: ϕ = soil porosity (mm^3/mm^3),
S_I = average initial saturation of the frozen soil layer at the start of infiltration (mm^3/mm^3)
z_f = depth of frozen soil at the start of infiltration (as measured from the ground surface)

The primary purpose for this constraint to index when a soil becomes unfrozen and the parametric expression no longer applies. That is, once infiltrating meltwater has satisfied the water storage potential of the frozen depth, a soil behaves as an unfrozen soil in respect to infiltration.

Note: usually a frozen soil is not normally saturated throughout its wetted depth by infiltrating meltwater during snow ablation because ice blocks the inter-connectivity of soil pores and the wetting front advances ahead of an a overlying layer of higher moisture. Thus, unless vertical movement is impeded completely, the average moisture content (water + ice) of the wetted zone seldom reaches W_{sP}. Gray et al. (1986) found that the average available storage potential of frozen mineral agricultural soils of the Canadian Prairies is approximately equal to 60% of the air-filled pore space at the start of infiltration.



List of Symbols

INF = infiltration estimate for the current time step

∑INF = cumulative infiltration estimate

U = the portion of the landscape unit that exhibits an unlimited capacity to infiltrate

R = the portion of the landscape unit that exhibits a restricted capacity to infiltrate

SWE = snow water equivalent

S_I = soil saturation, average (40 cm)

S_o = saturation applied at the surface

T_I = initial soil temperature, average (40 cm)

W_{sp} = water holding capacity of soil

Figure 20 Flow chart of infiltration model.

Discussion, Conclusions, Recommendations

The results above demonstrate that cold regions hydrological processes can have profound and previously undocumented impacts on the calculation of surface water and energy fluxes in the Mackenzie Basin. Progress has been made in understanding and describing many of the processes in a physical manner, in evaluating the process descriptions and in developing operational algorithms for specific processes. Some coupling, and/or comparisons of process algorithms with standard land surface scheme calculations have been demonstrated. The observed multi-scale operation and horizontal interaction of some of these processes means that phenomena operating at very small scales can affect large-scale water and energy balances. The relative success in transposing hydrological process descriptions from one environment to another can be attributed to the strong physical basis of the descriptions.

The following algorithms of cold regions hydrological processes have been developed, examined with respect to their performance and undergone enhancements where appropriate:

(1) Blowing snow model – sublimation.
(2) Complex terrain blowing snow model – snow accumulation.
(3) Intercepted Snow Accumulation/Unloading/Sublimation – sublimation.
(4) Accumulation and ablation of shallow snowcover of open environments.
(5) Boreal forest snow-cover - ablation.
(6) Infiltration to frozen soils – operational algorithm.

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