3.6 Hydrological Processes In Cold Regions

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1. Objectives

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

(1) ablation of seasonal snowcovers;
(2) coupled heat and mass transfer in snow and underlying ground;
(3) wind transport of snow; and,

To develop physically-based algorithms that describe these processes using field measurements in boreal, alpine and arctic environments.

2. Progress and Collaborations

2.1 Field Work

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

(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 CAGES site, 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 characterised 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.

2.2 Modelling

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

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

2.3 Collaborations

Collaborations in field work, modelling and analysis with GEWEX investigators: Pomeroy, Marsh, Granger, Woo, Pietroniro, Schuepp; with Janowicz at DIAND, Whitehorse, Yukon; Essery at the Hadley Centre for Climate Prediction and Research, UK Meteorological Office and Fortin (INRS-eau).

3. Scientific Results

3.1 Ablation of Seasonal Snowcovers

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 1).

The variability of melt energy within a stand decreased with overall stand density. 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 for stands with 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 2 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).

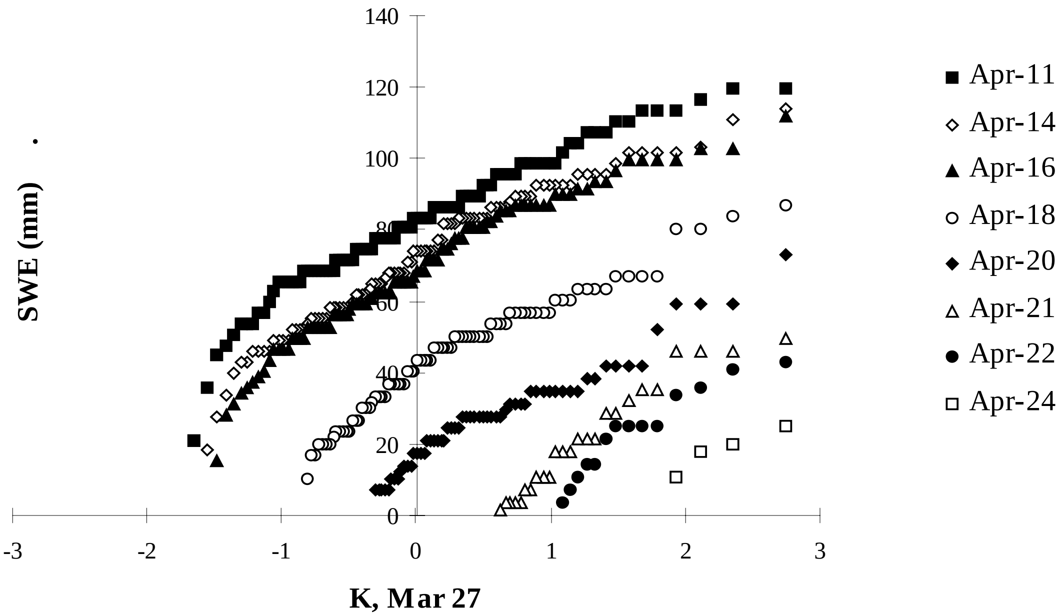

Figure 1 **Sequential distributions of snow water equivalent (SWE) during melt in a Pine Stand. K is the frequency factor for the log-normal distribution of SWE), the K for SWE=0 reflects snow covered area. Note that melt is greater for smaller SWE.** (after Faria et al., in press)

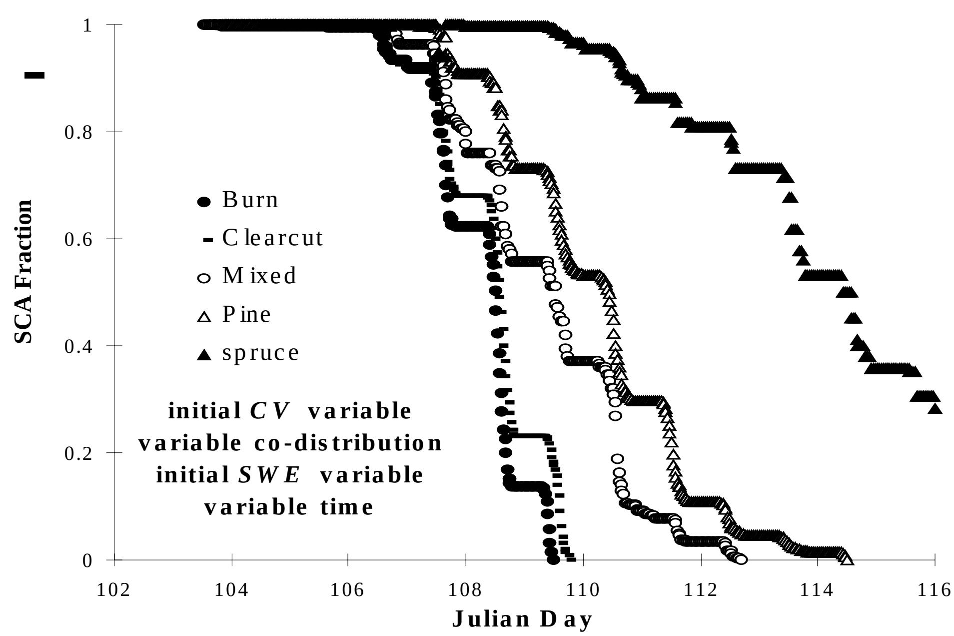

Figure 2 **Simulated snowcover depletion curves using measured mean melt rate at each site to calculate change in snow covered area with time as a function of the distribution of SWE and the covariance between SWE and melt rate.** (after Faria et al., in press)

3.2 Sublimation of Intercepted Snow

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 modification of CLASS's treatment of turbulent transfer and within-canopy ambient humidity were required to accommodate this nested control volume approach (Figure 3).

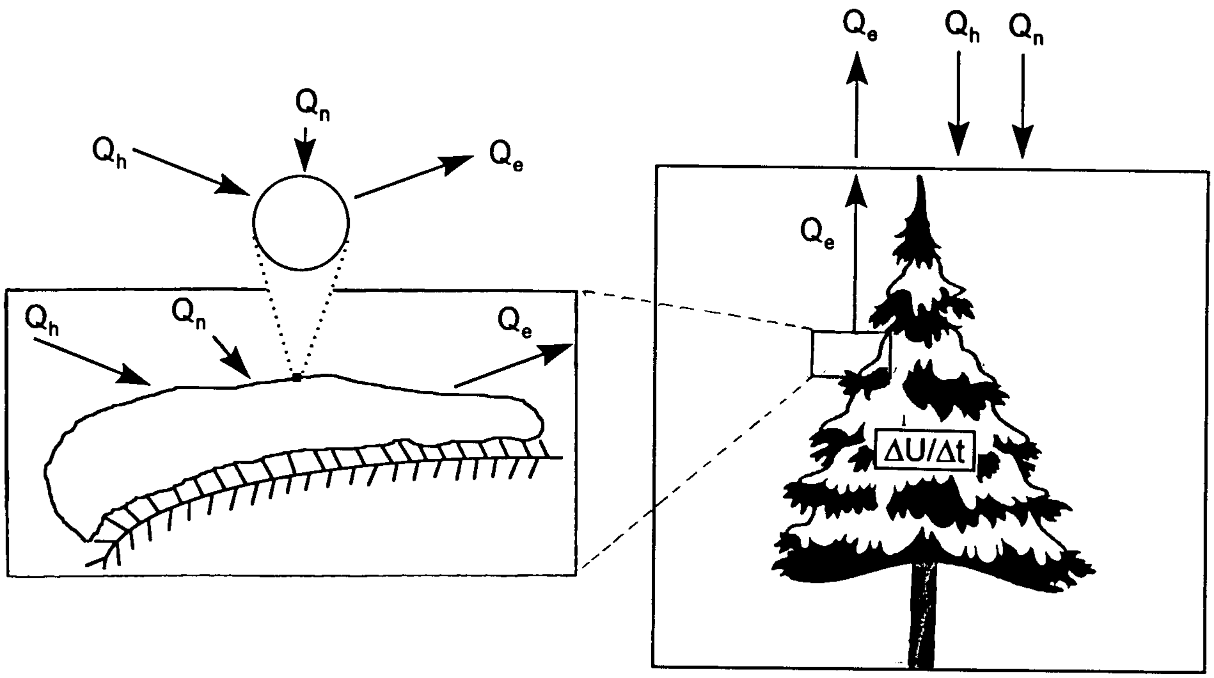

Figure 3 Nested control volumes for calculation of coupled mass and energy exchange between the atmosphere and intercepted snow. (after Parviainen and Pomeroy, submitted)

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 4). 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 vapour pressure, and 0.04 kg/m^2, respectively.

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.

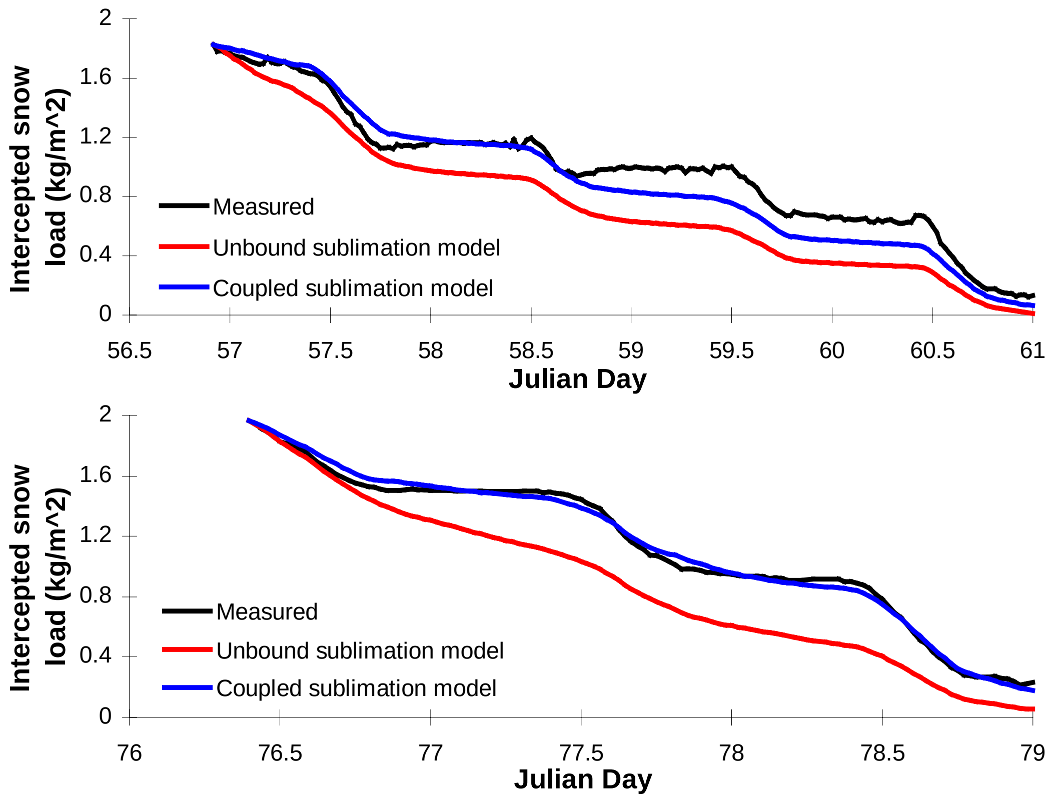

Figure 4 **Measured intercepted snow load (from suspended pine tree), and unbounded sublimation model and coupled land surface scheme sublimation model calculated intercepted load. Unbounded sublimation model driven using snowfall and within canopy measured wind speed, radiation, temperature and humidity. Coupled model driven with snowfall and above canopy reference wind speed, radiation, temperature and humidity.** (Pomeroy et al., 1998)

3.3 Blowing Snow

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). These fluxes are calculated as part of blowing snow two-phase particle transport models with provision for phase change based upon a particle-scale energy balance. Blowing snow models have normally been evaluated based upon their ability to reproduce diagnostic mass flux gradient measurements and regional-scale snow redistribution patterns and snow mass, e.g., Pomeroy and Li, submitted; Essery et al. (1999). Direct evidence has been obtained that large latent heat fluxes (40 to 60 W m^{-2}) that result in sublimation rates of 0.05 to 0.075 mm snow water equivalent $hour^{-1}$, are associated with mid-winter, high-latitude blowing snow events (Figure 5).

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, which can be predicted by standard bulk aerodynamic transfer equations, suggesting that blowing snow physics will have to be incorporated in land surface schemes and hydrological models in order to properly represent snow surface mass and energy exchange during blowing snow events (Pomeroy and Essery, 1999).

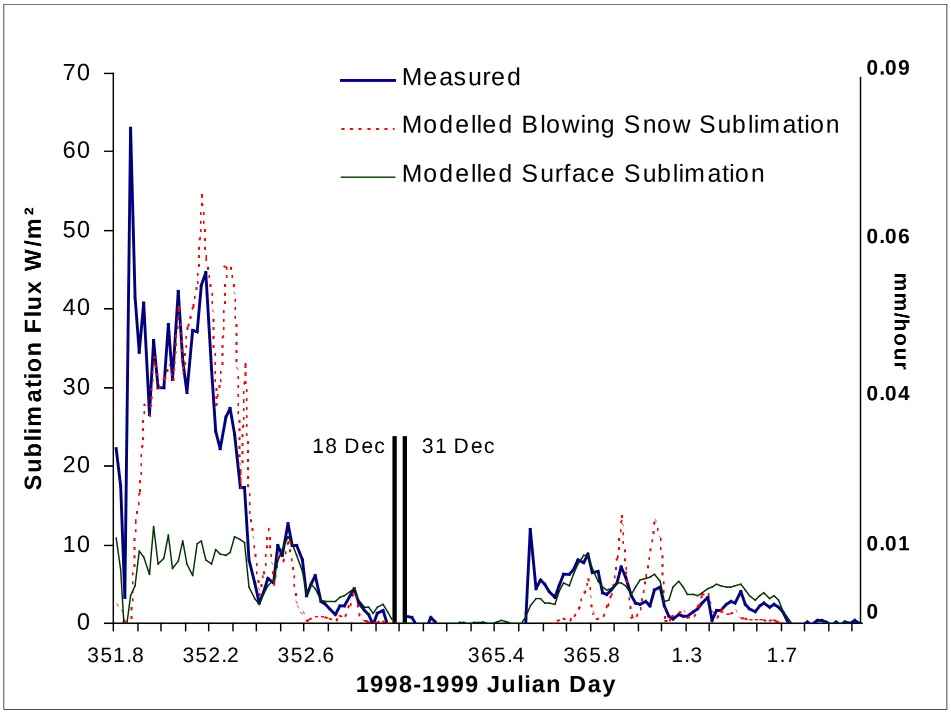

Figure 5 Measured sublimation flux, modelled blowing snow sublimation (PBSM) and modelled surface sublimation (bulk transfer) measured at over a level Prairie surface. (after Pomeroy and Essery, 1999)

3.4 Infiltration

Previous studies (Zhao and Gray, 1997; 1998; and 1999) have reported the development and testing of a general parametric correlation for estimating snowmelt infiltration into frozen soils. The expression relates cumulative infiltration, INF, to the soil surface saturation during melting, S_o, the total soil moisture saturation (water + ice), S_I, and temperature, T_I, at the start of snow ablation, and the infiltration opportunity time - the time that meltwater is available at the soil surface for infiltration, t, as:

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

in which C is a bulk coefficient that characterizes the effects on infiltration of differences between model and natural systems.

Up to now, the parametric expression (Equation 1) has been tested against field measurements of seasonal infiltration. During snow ablation in 1999, a series of field measurements were conducted at sites in the Prince Albert National Park near Waskesiu, SK and on the Kernen Research Farm at Saskatoon, SK to obtain data to verify estimates of infiltration by the expression over short time periods. The textures of the soils at the two sites are distinctly different; a sandy loam (43.0% sand, 11.0 clay) at Waskesiu and silty clay (4.4% sand, 49.0% clay) at Saskatoon. The sites were instrumented to provide information on air, snow and soil temperatures, net radiation, soil heat flux and profiles of changes in soil moisture (water + ice) over consecutive measurement dates. These data were used to derive estimates of snowmelt infiltration and infiltration time and to establish the initial and boundary conditions for the physically-based numerical model, HAWTS (Heat And Water Transport in Frozen Soils, Zhao et al., 1997). This simulation was used to derive the parametric equation (Zhao and Gray, 1999).

Figures 6 and 7 compare modeled and measured profiles of soil moisture (water + ice) at different infiltration opportunity times at the two sites. Figure 6 shows excellent agreement among profiles in the silty clay at the Kernen Farm and reasonable agreement (likely within measurement accuracy) between measured and modelled values in the sandy loam soil at Waskesiu. Similarly, there is good agreement among measured and modelled cumulative infiltration with a maximum difference of about 3.5 mm. These results suggest that the parametric expression (Equation 1) will give reasonable estimates of snowmelt infiltration.

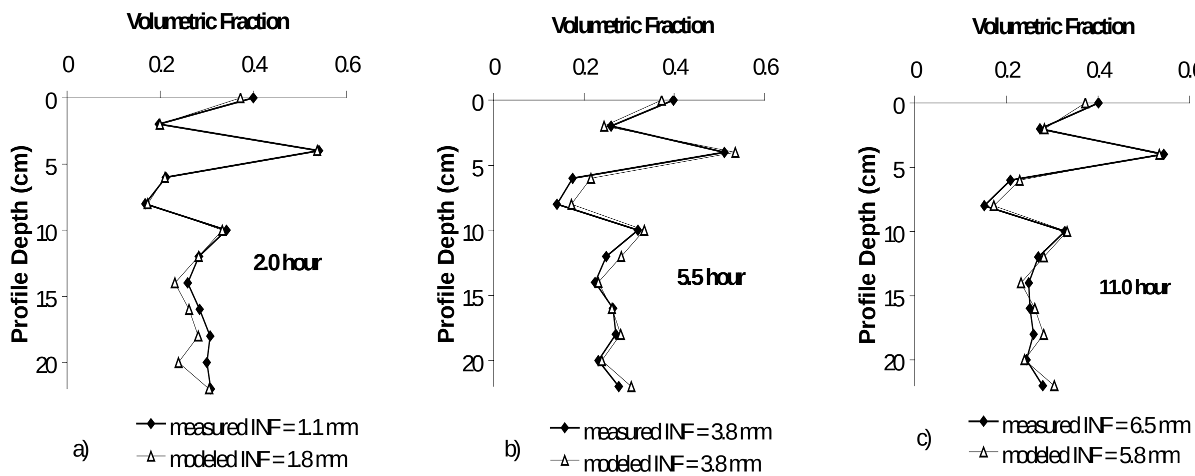

Figure 6 **Comparison of modeled and measured profiles of soil moisture (water + ice) into a frozen silty clay soil at the Kernen Farm after 2 h (Fig. 6a), 5.6 h (Fig. 6b) and 11.0 h (Fig.6c) of snowmelt infiltration simulation, silty clay soil. Simulation initiated at 1200 h March 17/99 and compared to measurements of soil moisture (water + ice) at: a) 1700 h March 17/99, b) 1630 h March 18/99 and c) 1100 h March 19/99.**

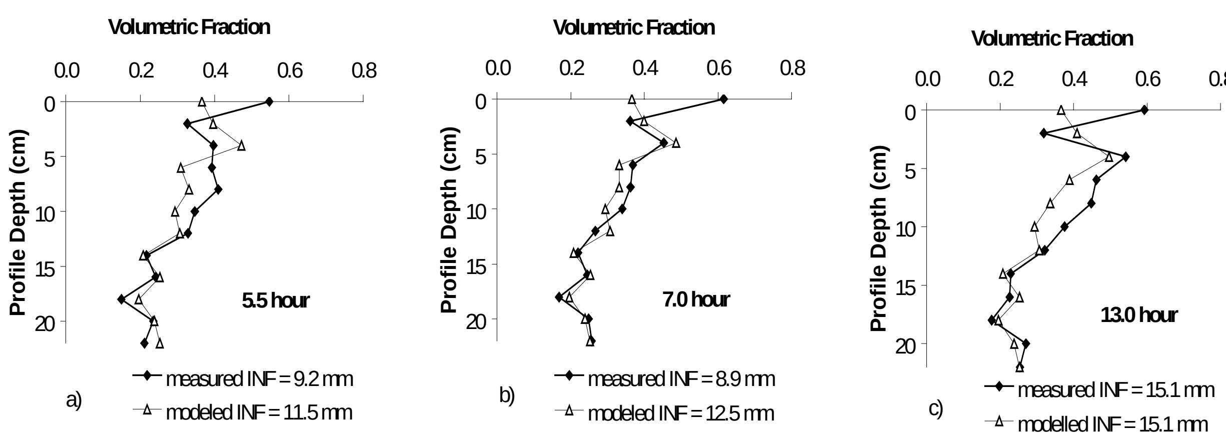

Figure 7 **Comparison of modeled and measured profiles of soil moisture (water + ice) into a frozen sandy loam soil at the Prince Albert Model Forest after 5.5 h (Fig. 7a), 7.0 h (Fig. 7b) and 13.0 h (Fig.7c) of snowmelt infiltration. Simulation initiated at 1200 h March 25/99 and compared to measurements of soil moisture (water + ice) at: a) 1730 h March 25/99, b) 0900 h March 26/99 and c) 1800 h March 26/99.**

As noted above, the coefficient C in Equation 1 characterizes the effects on infiltration of differences between model and natural systems. For example, the expression assumes surface saturation is constant, a uniform, and homogeneous soil, and the soil moisture and temperature throughout the soil profile at the start of infiltration are constant. These conditions are rarely found in nature. Zhao and Gray (1999) suggest representative values of C=1.0 - 1.3 for frozen sandy soils in a boreal forest and C=2.05 various fine-textured (sandy loam, loam, silty clay and clay) frozen Prairie soils.

The possibility that C may vary substantially is a serious limitation to the use of Equation 1 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. In these tests a linear variation in moisture with depth was assumed and the cumulative infiltration after 12 h calculated. This amount was normalized to the cumulative infiltration after 12 h assuming the soil moisture throughout the profile was uniform, the ratio is referred to herein as the infiltration ratio, IR. Figure 8 plots the results of these tests. The “x”-axis of this figure (Δ) represents the volume difference (expressed as a volume fraction) due to the distribution of soil moisture. That is, negative values of Δ represent a profile in which soil moisture content is increasing with depth and positive values of Δ represent a profile in which soil moisture content is decreasing with depth.

The data in Figure 8 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, with maximum values falling in the range from approximately 2.0 to 3.25, depending on the depth of the wetting front, for Δ=0.10. Conversely, wetter soils near the surface suppress infiltration, causing IR < 1. Tests have shown that the curves are independent of the other parameters, which comprise the parameteric equation (Equation 1).

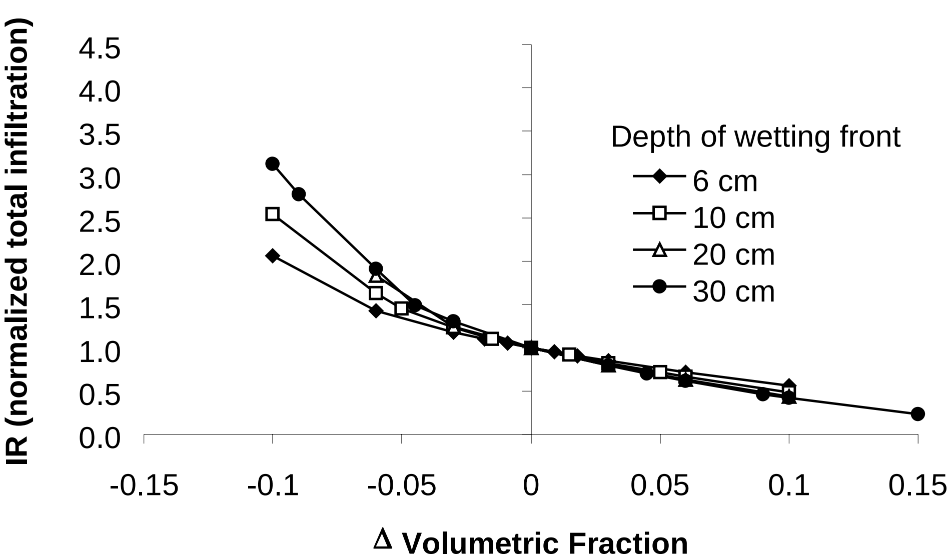

Figure 8 Variation in infiltration ratio with change in soil moisture (Δ) due to vertical gradient.

4. Summary

The results listed 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 describing many of the processes in a physical manner, evaluating the process descriptions and in developing operational algorithms for some of the processes. Some coupling, and/or comparison of process algorithms with standard land surface scheme calculations has 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.

5. Recent MAGS Publications/Presentations

Buttle, J.,I. Creed, and J.W. Pomeroy, 1999. Advances in Canadian Forest hydrology, 1995-1998. *Hydrological Processes*. (in press)

Elliott, J.A., B.M. Toth, R.J. Granger, and J.W. Pomeroy, 1998. Soil moisture storage in mature and replanted sub-humid boreal forest stands. *Canadian Journal of Soil Science*, **78**, 17-27.

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Faria, D.A., J.W. Pomeroy, and R.L.H. Essery, 1999. Forest canopy influences on snowcover depletion. *Hydrological Processes*. (submitted)

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Pomeroy, J.W. and E. Brun, 1999. Physical properties of snow. In: H.G. Jones, J.W. Pomeroy, D.A. Walker, and R.W. Hoham (Eds.), *Snow Ecology*, Cambridge University Press. (in press)

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