Chapter 5

Studies on Snow Redistribution by Wind and Forest, Snow-covered Area Depletion, and Frozen Soil Infiltration in Northern and Western Canada

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This chapter is dedicated to the memory of Don Gray, one of the original investigators in MAGS and a scientist of many contributions to cold regions hydrology, who passed away in January 2005, after a brief illness.

Abstract Important advances in our understanding of snow and frozen soil processes have been made, especially in regard to the transport and sublimation of blowing snow, interception and sublimation of snow in forest canopies, snow spatial distributions in complex environments, snowmelt in open environments and under forest canopies, advection of energy from bare ground to snow, snowcover depletion during melt, and heat and mass transfer during infiltration to unsaturated frozen mineral soils. These studies, conducted at the Division of Hydrology at the University of Saskatchewan, covered a range of northern environments including the tundra–taiga transition, the cordilleran sub-arctic, the southern boreal forest, and the northern prairie. Results from field research have led to the development and improvement of algorithms related to snow and infiltration processes, which have contributed to hydrologic and atmospheric models in the Mackenzie GEWEX Study.

1 Introduction

Storage, melt, infiltration, and runoff related to the seasonal snowcover are primary hydrologic events in most cold regions including the Mackenzie River Basin (MRB). Building on the 30-year tradition of cold regions hydrology research at the Division of Hydrology at the University of Saskatchewan, a research program was implemented to improve the understanding of the physical processes governing snow accumulation, ablation, and infiltration. These processes became the subject of extensive investiga-
tions in the Mackenzie GEWEX Study (MAGS). Led by Gray and Pomeroy from 1992 to 1999, field studies were conducted using a transect of instrumented research basins in different cold region environments, viz., the arctic–taiga transition, the cordilleran sub-arctic, the southern boreal forest, and the northern prairie. Physically-based algorithms were devised that describe these processes, and many of these algorithms have subsequently been incorporated in a range of hydrologic and atmospheric models.

This chapter presents the major findings pertaining to the investigation and modeling of several processes important to the water and energy cycles of cold environments, including interception and sublimation of forest snow, blowing snow transport and sublimation, snowmelt energetics and depletion of snow-covered area, and the infiltration of meltwater into frozen soils.

2 Methodology

Both field investigation and modeling were used in the study of snow and infiltration processes.

Intensive research on snow accumulation, ablation, and frozen soil infiltration were conducted at four locations, each representing a major northern environment:

1. Inuvik, Northwest Territories (arctic–taiga transition), e.g., Pomeroy et al. 1995
2. Wolf Creek, Yukon Territory (cordilleran subarctic), e.g., Pomeroy and Granger 1999
3. Waskesiu, Saskatchewan (southern boreal forest), e.g., Pomeroy et al. 1997a
4. Kerens Farm, Saskatchewan (northern prairie), e.g., Shook and Gray 1996

Field measurements employed standard meteorological observations as well as direct observations of sensible and latent heat fluxes over snow surfaces using eddy correlation systems, radiation, soil heat flux, blowing snow flux from an optoelectronic particle detector, intercepted snow mass measurements with a suspended, weighed full-size coniferous tree, and infiltration to frozen soils using twin-probe gamma attenuation devices, soil thermocouples, and time domain reflectometry.
Algorithms were developed for the following cold regions hydrologic processes. These algorithms were examined with respect to their performance and they have undergone enhancements where appropriate:

1. Canopy snow interception, unloading, and sublimation (Essery et al. 2003; Hedstrom and Pomeroy 1998; Parviainen and Pomeroy 2000; Pomeroy and Gray 1995; Pomeroy and Schmidt 1993; Pomeroy et al. 1998b)

2. Blowing snow model – threshold condition, scaling, and sublimation (Li and Pomeroy 1997a, b; Pomeroy and Li 2000; Pomeroy et al. 1998a)

3. Complex terrain blowing snow, redistribution, and sublimation (Essery 2001; Essery et al. 1999; Pomeroy et al. 1997b)

4. Snow ablation, open environment snowmelt, snowcover depletion, and small-scale advection to patchy snow (Faria et al. 2000; Pomeroy and Granger 1997; Pomeroy et al. 1998a, 2003; Shook and Gray 1996)

5. Infiltration to frozen soils – fully coupled mass and energy balance and operational algorithm (Gray et al. 2001; Zhao and Gray 1997; Zhao et al. 1997).

3 RESULTS

3.1 Interception, Unloading and Sublimation of Canopy Snow

Field results from several winters of observation of snow accumulation under forest canopies, snowfall, and the load of snow on a series of cut, weighed, suspended trees show that leaf area, canopy closure, species type, time since snowfall, snowfall amount, and existing snow load control the efficiency of snow interception (Hedstrom and Pomeroy 1998; Pomeroy and Gray 1995). A physically-based algorithm describing these results was developed from observations using a weighed suspended tree (see Fig. 2 in Woo and Rouse 2007). The sensitivity of this algorithm to winter leaf area index, air temperature, snowfall and initial canopy snow load is shown in Fig. 1.

Physically-based equations describing snow interception (Hedstrom and Pomeroy 1998) and sublimation processes (Pomeroy et al. 1998b) were coupled and applied to calculate canopy intercepted snow load using a fractal scaling technique (Pomeroy and Schmidt 1993) to provide a snow-covered forest boundary condition for the one-dimensional Canadian Land Surface Scheme, CLASS (Verseghy et al. 1993). Parviainen and Pomeroy
Pomeroy et al. (2000) found that CLASS could be modified to provide an appropriate treatment of turbulent transfer and within-canopy ambient humidity to accommodate this nested control volume approach. 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 (Fig. 2). A complete model was tested in Russia and confirmed with long term data (Gelfan et al. 2004). The model has been incorporated into the Hadley Centre land surface scheme and compared well to a range of field measurements from Canada and the United States (Essery et al. 2003). Pomeroy et al. (2002) developed a parametric form of this model to predict snow accumulation in northern forests as a function of accumulation in small clearings or cumulative snowfall.

### 3.2 Blowing Snow

Sublimation fluxes during blowing snow have been estimated to return 10–50% of seasonal snowfall to the atmosphere as vapor in North American prairie and arctic environments (Essery et al. 1999; Pomeroy and Gray 1995; Pomeroy et al. 1997b). These fluxes are calculated as part of blowing snow two-phase transport models with provision for phase change based upon a particle-scale energy balance driven by measured wind speed, air temperature, and humidity (Pomeroy and Li 2000; Pomeroy et al. 1993) and a snow transport threshold algorithm that uses air temperature to index snow cohesion (Li and Pomeroy 1997a). The probability of
occurrence of blowing snow over time or space (for uniform terrain) follows a cumulative normal distribution that is controlled by snow temperature, snow age, vegetation exposure, and occurrence of melt or rain (Li and Pomeroy 1997b). An algorithm describing blowing snow probability provides a means to scale blowing snow fluxes from point to large areal averages in a computationally simple manner (Pomeroy and Li 2000). An example of the model operation for tundra surfaces at Trail Valley Creek is shown in Fig. 2. The model was tested extensively in the prairie environment as well and found to redistribute snow appropriately between land cover classes (Pomeroy et al. 1998a).

Landscape classifications from a LANDSAT image, a complex terrain workflow model, snowmelt, and blowing snow process routines can be used to determine blowing snow fluxes over irregular arctic land surfaces. The resulting runs with a distributed blowing snow model represented the distribution of SWE in test basins and matched basin snow accumulation within 6% (Pomeroy et al. 1997b). Sublimation losses were small for the subarctic basin, about 21% over the arctic basin and 30% from tundra surfaces (Pomeroy and Marsh 1997). Subsequent tests with a more physically-based distributed blowing snow model show that the arctic tundra comprises a variety of blowing snow flow zones that are largely controlled by vegetation cover and by topographically induced zones of convergence and divergence of windflow. Results that included sublimation as modeled
by Pomeroy and Li (2000) produced spatial distributions of snow accumulation that were close to the values obtained by extensive snow surveys (Essery et al. 1999). However, results with suppressed blowing snow sublimation redistributed too much snow to the treeline, where modeled accumulations were much greater than observed. This provides a regional mass balance confirmation of significant sublimation from blowing snow on arctic tundra and calls into question models that produce minimal sublimation from blowing snow in this environment. Later work developed scaling relationships from this model (Essery 2001) and examined the influence of shrub tundra coverage on blowing snow fluxes (Essery and Pomeroy 2004). Figure 3 provides an example of the mapped SWE distribution simulated with the blowing snow model, for an arctic domain centered about Trail Valley Creek.

**Fig. 3.** Mapped distribution of late winter snow accumulation (mm SWE) in the Trail Valley Creek domain; simulation produced with the Distributed Blowing Snow Model (DBSM), a version of PBSM coupled to the MS3DJH/3R complex terrain boundary-layer model (after Essery et al. 1999)

Blowing snow models such as the Prairie Blowing Snow Model (PBSM) have normally been evaluated based upon their ability to reproduce diagnostic mass flux gradient measurements and regional-scale snow redistribution patterns and snow mass. Direct evidence was obtained in MAGS that large latent heat fluxes (40–60 W m$^{-2}$) that result in sublima-
tion rates of 0.05–0.075 mm-SWE hour⁻¹ are associated with mid-winter, high-latitude blowing snow events (Fig. 4). For events with wind speeds above the threshold level for snow transport, these fluxes are within the range of those predicted by Pomeroy and Li (2000). The fluxes are well in excess of those which can be predicted by standard bulk aerodynamic transfer equations, suggesting that blowing snow physics needs to be addressed by land surface schemes and hydrologic models in order to properly represent snow surface mass and energy exchange in open environments (Pomeroy and Essery 1999).

Fig. 4. Measured sublimation flux, modeled blowing snow sublimation (PBSM) and modeled surface sublimation (bulk transfer) measured at a level Prairie surface (after Pomeroy and Essery 1999)

### 3.3 Ablation of Seasonal Snow-covers

Advection of energy from bare ground to patchy snow was recognized to have an important role in the energy equation for snowmelt in open environments. Shook (1995) applied the advection expressions of Weisman (1977) in small-scale gridded model to a synthetic snowcover; grid cells as they became snow-free became sources of advected energy to the remaining snow-covered cells. The synthetic snowcover was generated using the
Fractal Sum of Pulses Method and had the same statistical properties as real snowcovers (Shook and Gray 1997a). Shook and Gray (1997b) showed that advection was important for late spring snowmelt under strong insolation as is typical in the northern prairies and arctic. This detailed model is the basis behind the simplified advection efficiency scheme of Marsh and Pomeroy (1996) and Marsh et al. (1997).

Because snow depth is self-similar and fractal, the frequency distribution of snow water equivalent (SWE) can be simulated using a 2-parameter log-normal density function (Shook and Gray 1996). By applying an even melt rate to this distribution the snow covered area (SCA) can be calculated during snow depletion (Shook and Gray 1997c). The more variable the initial SWE, the more rapid the SCA depletion, other factors held constant (Fig. 5).

![Graph](image)

**Fig. 5.** Snow-covered area depletion curves modeled by applying a constant, uniform melt rate to snow covers with SWE that have a log-normal distribution. Fractional snow cover is plotted against time from an initial SWE of 130 mm for various coefficients of variation (CV) of SWE.

Pomeroy et al. (1998a) provided the statistical parameters needed to calculate snow depletion for various environments and showed how this might be incorporated in melt models. When this approach was tested in complex terrain, it was shown that variable melt energy needs be addressed, even in
open environments, but this could be done by landscape stratification (Pomeroy et al. 2003).

The influence of forest canopy cover and variable melt energetics on depletion of snowcover in the boreal forest was then investigated. The results can be distinguished between variability within the forest stand and that between forest stands. Between forest stands, Pomeroy and Granger (1997) found that melt energy was much lower as stand density increased. The radiation model developed by Pomeroy and Dion (1996) accounts for the shortwave component of these melt differences. An examination of the shrub tundra environment led to extension of these concepts to shrub terrain that covers much of the sub-arctic (Pomeroy et al. 2006). Within stands, Faria et al. (2000) found the frequency distribution of SWE under boreal canopies fit a log-normal distribution, with the most dense stand displaying the most variable SWE prior to melt. Within stands, snowmelt energy below the canopies was found to be spatially heterogeneous and inversely correlated to SWE (Fig. 6).

![Graph](image)

**Fig. 6.** 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. 2000)
The variability of melt energy within a stand decreased with overall stand density. Within-stand covariance between the spatial distributions of SWE and melt energy promoted an earlier depletion of snowcover than if melt energy were uniform (Pomeroy et al. 2001). This covariance was largest for the most heterogeneous stands (usually medium density). Stand scale variability in mean SWE and mean melt energy resulted in the most 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 SCA depletion (Faria et al. 2000). Figure 7 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. 2000), and the between-stand variation is consistent with the findings of Pomeroy and Granger (1997).

![Simulated snowcover depletion curves](image)

**Fig. 7.** 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. 2000)
3.4 Infiltration to Frozen Soils

Infiltration into frozen ground involves simultaneous coupled heat and mass transfers with phase changes, such that the infiltrating water conveys heat transfer into the ground and modifies the soil temperature regime. Field measurements and model simulations (Zhao et al. 1997) demonstrate that both the infiltration rate and the surface heat transfer rate (conduction) in a frozen soil decrease with time following the application of meltwater to the surface (Fig. 8). Zhao et al. (1997) separate these variations into a transient regime and a quasi-steady-state regime. The transient regime follows immediately the application of water on the surface and during this period the infiltration rate and the heat transfer rate decrease rapidly. The quasi-steady-state regime occurs when the changes in the infiltration rate and the heat transfer rate with time are relatively small. The duration of the transient period is usually short (a few hours) and the energy used to increase the soil temperature is largely supplied by heat conduction at the surface (i.e., high heat transfer rate at the surface). 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 (i.e., low heat transfer rate at the surface). Zhao et al.

![Fig. 8. Variations in infiltration rate (dINF/dt) and surface heat flux rate (dQ/dt) with time during snowmelt infiltration into a frozen silty clay soil (after Zhao et al. 1997)](image-url)
1997) estimate that as much as 90% of the latent heat released by the refreezing of meltwater is conducted deeper in the soil where it used for melting and increasing the soil temperature.

Zhao and Gray (1997, 1999) and Gray et al. (2001) reported the development and testing of a general parametric correlation for estimating snowmelt infiltration into frozen soils. Cumulative infiltration \(\text{INF}\) is expressed through the following relationship with the soil surface saturation \(S_o\) during melting, the total soil moisture saturation (water and ice) \(S_i\), the temperature \(T_i\) in K at the start of snow ablation, and the infiltration opportunity time \(t\) (i.e., the time that meltwater is available at the soil surface for infiltration):

\[
\text{INF} = C S_o^{2.92} (1 - S_i)^{0.64} \left[ \frac{273.15 - T_i}{273.15} \right]^{-0.45} t^{0.44}
\]

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

Figure 9 compares modeled and measured profiles of soil moisture (water and ice) at different infiltration opportunity times in the boreal forest. There is reasonable agreement (likely within measurement accuracy) between measured and modeled values in the sandy loam soil at Waskesiu, and similarly good agreement among measured and modeled cumulative infiltration with a maximum difference of about 3.5 mm, indicating that Eq. (1) gives reasonable estimates of snowmelt infiltration.

The coefficient \(C\) in Eq. (1) characterizes the effects on infiltration of differences between model and natural systems. For example, the expression assumes that surface saturation is constant, the soil is uniform and homogeneous, 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 \(C = 2.05\) for various fine-textured (sandy loam, loam, silty clay, and clay) frozen Prairie soils.

4 Conclusions

The results reported demonstrate that cold regions hydrologic processes can have profound and previously undocumented impacts on the calculation of surface water and energy fluxes of the land surface environments found in the MRB. Progress has been made in describing many of the pro-
Fig. 9. Comparison of modeled and measured profiles of soil moisture (water + ice) into a frozen sandy loam soil in Prince Albert National Park after (a) 5.5 h, (b) 7.0 h, and (c) 13.0 h of snowmelt infiltration. Simulation initiated at 1200 h March 25, 1999 and compared to measurements of soil moisture (water + ice) at: (a) 1730 h March 25, (b) 0900 h March 26, and (c) 1800 h March 26.
cesses 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 hydrologic process descriptions from one environment to another can be attributed to the strong physical basis of the descriptions employed.

References


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