HYDROLOGICAL PROCESSES
Hydrol. Process. **14**, 1591–1611 (2000)

Snow, frozen soils and permafrost hydrology in Canada, 1995–1998

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Abstract:

This paper provides an overview of Canadian research on snow, frozen soils and permafrost hydrology during the years 1995–98. There were significant advances in the understanding of processes and the development of models of snow accumulation and melt, including the relocation of snow by wind, snow interception in forest canopies, sublimation and energy balance during snowmelt. A major aspect was the development of physically based predictive techniques that account for the effects of heterogeneous topography, vegetation and snow properties, and complex boundary-layer development on snow accumulation, evaporation, melt and runoff. Another advancement is in the linkage of physical snow processes with chemical models to better describe ion accumulation and elution from snow. Snow ecology has shown the interactions in nutrient cycles that involve snow. Frozen ground research has resulted in significantly improved models of frozen soil infiltration, based on both field observations and thermodynamic principles. Research in permafrost regions includes the exfiltration of groundwater in the seasonally thawed zone and the occurrence of perennial springs discharged from below the permafrost. Groundwater discharge is important to features such as icings and the occurrence of wetlands in a polar desert. Processes governing runoff generation on hillslopes have been examined, both in continuous and discontinuous permafrost zones. In terms of future research directions, consideration should be given to continued intensive field studies of cold region hydrological processes and the incorporation of these processes into aquatic chemistry and hydrological models and land surface schemes used in atmospheric models. A better understanding of the role of hydrological boundaries in affecting the rates of processes is needed. The question of scaling processes up from the small scale at which they are relatively well understood, to the larger scales necessary to address global environmental concerns also should be addressed.

KEY WORDS snow; snowcover; snowmelt; infiltration; frozen ground; frost; permafrost; Canada

INTRODUCTION

The large area of Canada that is cold during important parts of the hydrological cycle stimulates practical and scientific research interests in the hydrology of ice, snow, frozen soil and permafrost. Compilations of Canadian research on these topics have been prompted in the past by the Canadian National Committee for the International Hydrological Decade (CNC-IHD, 1970; Gray, 1970), National Research Council of Canada (e.g. Brown, 1969; Gray and Male, 1981; French, 1982; Kry, 1987) and the national Hydrology Research Institute, Environment Canada (Prowse and Ommanney, 1990). More recently, Canadian snow and cold region hydrologists have been active participants in international meetings such as symposia of the International Commission on Snow and Ice, International Association of Hydrological Sciences (e.g. Tonnessen *et al.*, 1995; Jones *et al.*, in press), the Northern Research Basin Symposia/Workshops (Northern

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Received 10 May 1999
Accepted 5 October 1999

Research Basins, 1997), joint meeting of the Canadian Geophysical Union Hydrology Section, and the Eastern and Western Snow Conferences hosted by Canada (Albert *et al.*, 1998), the International Conference on Snow Hydrology (Hardy *et al.*, 1999) and in national conferences such as the Annual Meetings of the Canadian Geophysical Union. These meetings and resulting special issues of hydrological journals provide the vehicles for the communication of Canadian work. This report concerns itself with those snow and cold region hydrological contributions presented and prepared in the period 1995–98. Ice-related, glaciological and remote sensing studies are summarized in other reports in this special issue (Beltaos, 2000, this issue; Munro, 2000, this issue; Pietroniro and Leconte, 2000, this issue).

Canada has always maintained strength in field investigations of snow and cold region processes, with a concentration of research on the scale of experimental plots to small drainage basins. By the mid-point of this decade Canadian snow and frozen ground hydrologists had identified several critical questions that they felt required focused research to address. Some questions were fundamental, illustrating the many research opportunities yet available in this field. The questions covered topics of snow, frozen soil and permafrost hydrology. Process hydrology was identified as an area requiring fundamental improvement, with respect to understanding snow interception, sublimation (including feedbacks), chemical transformation upon sublimation, gas transmission, melt energetics, influence of melt heterogeneity upon chemical fractionation, advection of energy during melt, thermodynamics of infiltration to frozen soils, and runoff mechanisms over permafrost hillslopes. The ability to scale processes (both up and down) was found to be generally lacking, with particular needs identified for blowing snow redistribution and sublimation, snow interception, snow chemistry, snow covered area depletion, snowmelt energetics and ablation, effects of varying aspect on permafrost hillslope hydrology and runoff over permafrost and seasonally frozen soils. Modelling of snow and frozen ground processes was identified as an area requiring improvement in Canada, as recent advances in process understanding had not yet been incorporated in Canadian hydrological and atmospheric models and the capacity to model the water balance of northern Canada in a physically realistic manner was extremely weak. Finally, interactions amongst snow, ice, frozen ground and the environment reflect considerations that are increasingly prominent and led to concerns about the importance of biogeochemical cycling to snow ecosystems, greenhouse gas transmission through snow, maintenance of wetlands by exfiltration from permafrost and the cumulative effects on runoff generation of various snow and frozen ground processes in the North. These questions are important for determining water balances, peak runoff, chemical transport and atmospheric exchanges in many regions of Canada. They contribute directly to issues of recognized national importance, such as preserving Arctic and Boreal environments, promoting sustainable agricultural and forestry production, and anticipating the impacts of climate and land use change on water resources, and to international issues, such as sustainable development of the Arctic, a better understanding of the role of the cryosphere in controlling climate and in evaluating continental-scale water and energy cycles.

The response to the questions posed in mid-decade has resulted in considerable increase in our knowledge of snow accumulation, redistribution, melt, frozen soil infiltration, seasonal frost and permafrost hydrology. Field studies were often followed by numerical modelling of the processes, using a physically based approach or in combination with empiricism. This bottom-up approach of developing models in conjunction with field results complements the conceptual modelling approach used in the construction of macroscale hydrological and land surface schemes. The advent of interdisciplinary projects such as CRYSYS (use of the cryospheric system to monitor global change in Canada) and Global Energy and Water Cycle Experiment (GEWEX; Rouse, 2000, this issue) has assembled the process researchers and the macroscale modellers from various fields, offering collaborative forums to which snow and frost hydrologists can contribute.

SNOW ACCUMULATION

Research on snow accumulation in both forested (deciduous and evergreen) and open environments (agricultural, alpine and arctic) has emphasized processes of snow redistribution, spatial patterns of snow

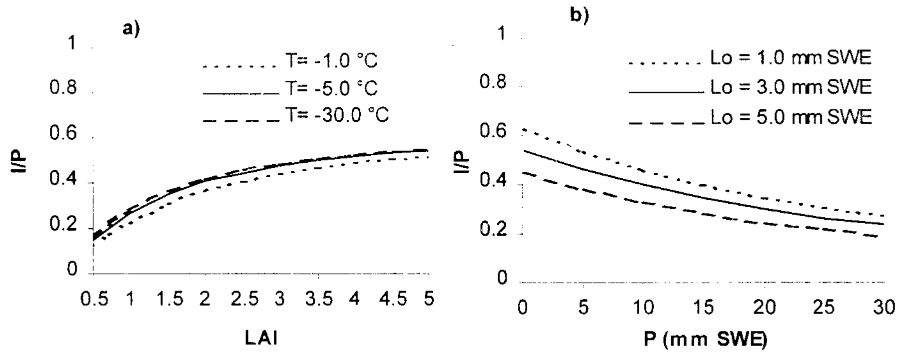

Figure 1. Modelled interception efficiency (snow interception/snowfall) as a function of (a) winter leaf-area index and air temperature, and (b) snowfall and initial canopy snow load (L_o). After Hedstrom and Pomeroy (1998)

accumulation over heterogeneous surfaces and concomitant sublimation fluxes that occur during accumulation periods. Detailed reviews on advances in snow accumulation and melt hydrology were provided by the book *Snow Accumulation, Redistribution and Management* (Pomeroy and Gray, 1995). A perspective on snow mass and energy exchange between atmosphere and surface, with an emphasis on snow accumulation in Canada, is found in a chapter by Pomeroy and Goodison (1997) in the *Surface Climates of Canada* and a review by Marsh (1999). The following discussion focuses on the physical processes of snow accumulation in forested and open environments.

Forested environments

Interception and sublimation of snow strongly control snow accumulation in forests. Increases of 30–45% in seasonal snow accumulation have been measured after the removal of evergreen forest cover by clear-cutting at sites across Canada (Pomeroy and Gray, 1995; Pomeroy *et al.*, 1997a). Forest canopies also influence the variability of snow water equivalent (SWE); Pomeroy *et al.* (1998a) found that coefficients of variation of SWE generally increase with increasing evergreen canopy density in the boreal forest, but Faria (1998) found the highest SWE variability under canopies of medium density and in mixed-wood forests.

A modification of the McNay *et al.* (1988) forest snow depth equation by Pomeroy and Gray (1995) enabled the estimation of the SWE accumulation as a linear function of evergreen canopy coverage and above-canopy snowfall. Subsequent research in the boreal forest yielded a physically based method to estimate snow interception as a function of leaf-area index, evergreen canopy coverage, tree species, initial canopy snow load, air temperature, wind speed, time since snowfall and snowfall amount (Hedstrom, 1998; Hedstrom and Pomeroy, 1998). An example of the effects of leaf-area index (LAI), initial snow load (L_o), snowfall amount (P) and air temperature (T) on interception efficiency (I/P) is presented in Figure 1. This example demonstrates that interception efficiency is strongly controlled by leaf area and snowfall, and that initial snow load in the canopy should be considered in calculating subsequent interception. Snowfall interception was shown to be much larger than rainfall interception and much greater than the parameterizations of snow interception presently used by land surface schemes (Pomeroy *et al.*, 1998a).

Sublimation reduces the amount of intercepted snow that is available for unloading to the ground surface. Pomeroy and Gray (1995) applied fractal geometry to parameterize snow surface area and used equations describing snow particle thermodynamics, turbulent and radiative exchange to calculate canopy snow sublimation. Because intercepted snow does not significantly increase the albedo of evergreen canopies (Pomeroy and Dion, 1996), mid-winter net radiation is positive and the snow-covered canopy provides a source of water vapour and a sink of sensible heat (Harding and Pomeroy, 1996). Pomeroy *et al.* (1998b) used the snow interception algorithms of Hedstrom and Pomeroy (1998) and the snow sublimation algorithms of Pomeroy and Gray (1995) to calculate canopy snow mass balance and surface snow accumulation;

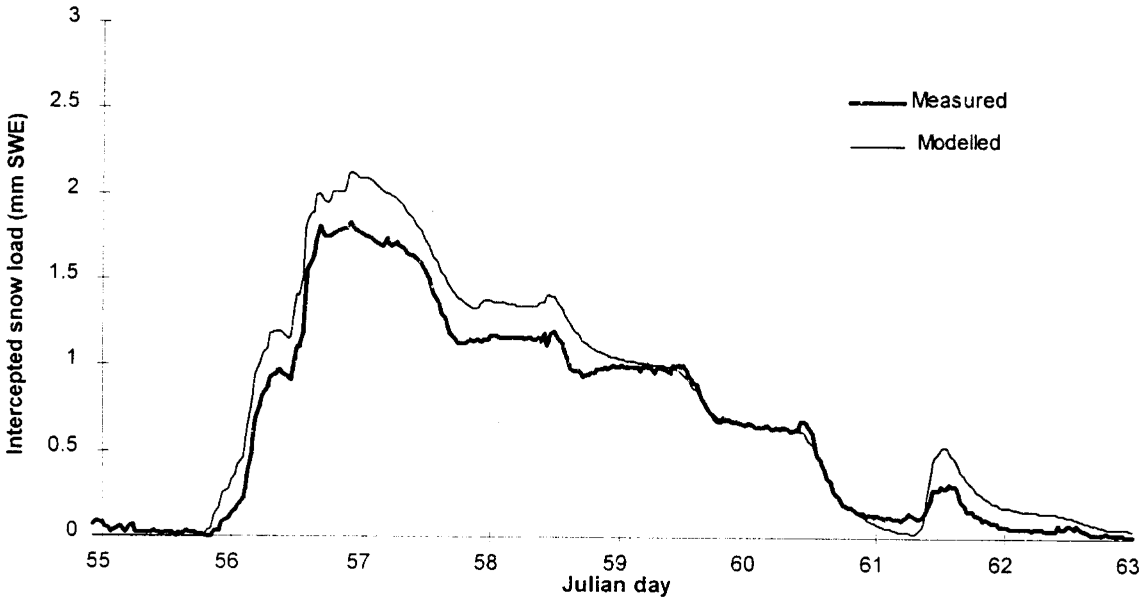

Figure 2. Modelled and simulated intercepted snow load in a pine forest: change in intercepted snow load is due to precipitation, unloading and sublimation. Measured snow load is derived from the mass of snow weighed on a suspended pine tree, scaled to areal snow water equivalent by comparative measurements of above canopy snowfall and below canopy snow accumulation. After Pomeroy *et al.* (1998b)

modelled values compared well with measurements and suggested that sublimation can ablate up to 5 mm SWE per day in the southern boreal forest. A canopy snow load accumulation and sublimation sequence in a boreal forest, measured using a weighed pine tree and modelled using Pomeroy *et al.*'s (1998b) interception/sublimation model, is shown in Figure 2 and illustrates the relatively rapid ablation of forest snow due to sublimation.

Open environments

In open environments, blowing snow can redistribute snow cover from areas of greater wind exposure to sheltered sites, with sublimation of blowing snow particles causing ablation and redistribution of the snow cover. Woo (1998) discussed scaling of snow cover information in the Arctic noting that at the microscale detailed mapping of SWE is required, at the mesoscale SWE patterns can be mapped and inferred from terrain units, and that data must be aggregated based on landscape units for large-scale applications. Lapen and Martz (1996) in a study of snow depth and topography in a prairie environment demonstrated that simple relationships between snow accumulation and topographic slope position descriptors are not apparent. Shook and Gray (1996) presented measurements of prairie SWE which show that the distribution of SWE is fractal at small scales, becoming random at scales beyond the 'limiting length' of about 30 m on level terrain and much longer on rolling terrain (at least the length of terrain features such as broad valleys). The implication of the fractal-limiting length is that snow survey transects should be at least this length to ensure an appropriate measure of standard deviation of SWE and that subgrid SWE at scales smaller than the length should be considered fractal, not random.

Pomeroy and Gray (1995) detailed the development and operation of the Prairie Blowing Snow Model (PBSM) for calculating snow transport and sublimation on agricultural fields in western Canada. Seasonal redistribution was estimated to remove up to 75% of snow cover from summer-fallow fields with fetches in excess of 4 km in the southern Canadian Prairies. Seasonal snow transport and sublimation calculated for

15 locations in western Canada showed that sublimation consumed from 15% to 40% of seasonal snowfall. A monthly climatological blowing snow model, based on the PBSM results, was applied to complex terrain in the western Arctic by Pomeroy *et al.* (1997b) and predicted sublimation losses of 28% of seasonal snowfall from tundra surfaces. This climatological model was verified in the Arctic by landscape-stratified measurements of snow accumulation and corrected snowfall measurements.

Li and Pomeroy (1997a) examined the variation of the threshold wind speed for the initiation of blowing snow transport and its relation to standard weather variables measured at 15 stations in the Prairie Provinces over six seasons. They found that the blowing snow transport threshold increased with air temperature (> -27°C) and the occurrence of melt or freezing rain, but diminished during snowfall, and produced an empirical expression to describe this. The probability of blowing snow occurrence was similarly investigated and found to fit a cumulative normal distribution that was controlled by air temperature and snow age (Li and Pomeroy, 1997b).

Pomeroy and Li (1997) added transport threshold, blowing snow probability, mid-winter melt and improved vegetation roughness routines to the PBSM and applied it to stations in western and central Canada. For Calgary, Regina, Prince Albert and Portage la Prairie they concluded that:

1. sublimation and transport ablate 12–33% and 3–16%, respectively, of annual snowfall over the winter;
2. the temporal pattern of snow accumulation is modelled correctly by PBSM although there are cumulative errors in the seasonal snow accumulation;
3. calculations of ablation due to mid-winter melt alone would lead to substantial error (16–49% of snowfall) in the SWE estimated for the end of the winter.

Pomeroy *et al.* (1999a) tested the modified PBSM against eddy correlation measurements of latent heat flux over a prairie blowing snow storm and found reasonably good agreement, providing a direct confirmation of the predicted sublimation fluxes for moderate blowing snow events.

At Bad Lake, Saskatchewan, comparison to snow surveys suggested that the PBSM provides good representation of the seasonal regime of snow accumulation over fallow, grain stubble fields and shrub-filled valleys. Sublimation and transport losses from fallow fields were 24 and 15% of seasonal snowfall and transport into shrub-filled valleys increased SWE by an additional 50 to 150% of seasonal snowfall (Pomeroy *et al.*, 1998a).

Essery *et al.* (1999) coupled a version of PBSM to a complex terrain wind-flow model, applied it to a digital elevation and vegetation cover grid of a rolling landscape in the western Arctic and tested it with distributed snow survey measurements of SWE. This complex-terrain blowing snow model predicted accumulation patterns and statistical distributions with a grid resolution of 80×80 m over a domain of 12×14 km (Figure 3); the patterns and distributions agreed well with field measurements from landscape-based snow surveys. Application of the model demonstrated that:

1. zones of blowing snow convergence and divergence exert important controls on SWE;
2. blowing snow is redistributed between catchments;
3. vegetation in the mixed shrub–tundra and open tundra strongly controls the evolution of SWE over the winter.

The complex terrain wind-flow model used by Essery *et al.* cannot be applied to mountainous terrain. For such terrain, improvements in wind-flow models and coupling with blowing snow boundary layer models are required.

Dery and Taylor (1996) and Dery (1998) developed a blowing snow model that, in addition to the transport and sublimation physics of PBSM, considers thermodynamics and mixing in the lower boundary layer. This model parameterizes atmospheric exchange of water vapour and sensible heat with sublimating blowing snow. It predicted that the transport of blowing snow was mostly limited to only several metres above the surface because of particle sublimation during upward diffusion. In the fetch-dependent version of the model, Dery *et al.* (1998) suggested that over fetches that exceed 1 km, the lower boundary layer cools and

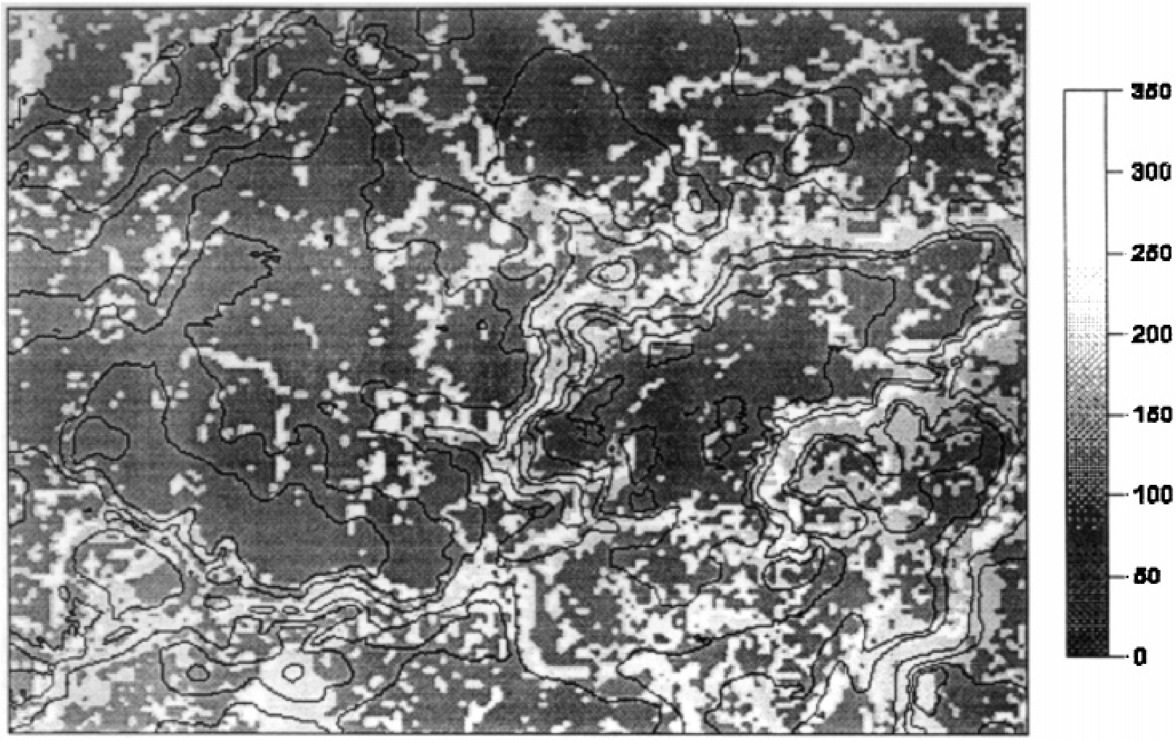

Figure 3. Mapped distribution of late winter snow accumulation (mm SWE) in the Trail Valley Creek domain (12 × 14 km), simulation produced with a version of PBSM (Pomeroy and Li, 1997) coupled to the Walmsley–Salmon–Taylor MS3DJH/3R complex-terrain boundary layer model. After Essery *et al.* (1999)

approaches saturation, suppressing blowing snow sublimation. Thus, sublimation losses for long fetches are reduced compared with the loss for short fetches, with maximum predicted to occur at fetches of about 1 km. Although the Dery *et al.* (1998) model has yet to be field verified, the difference between its conclusion and that derived from the PBSM (increasing sublimation loss with fetch distance) has important implications to scaling considerations in snow hydrology.

Spatial distribution, snow management and modelling

Studies continued in the Prairie Provinces on blowing snow management techniques and the impact of this trapping on spring soil moisture and crop yields. McConkey *et al.* (1997) found that tall (25 cm) wheat stubble increased spring soil water by 15 mm compared with conventional short stubble (14 cm tall). Steppuhn and Waddington (1996) investigated the effects of double rows of tall wheatgrass on snow accumulation and found spring recharge of water to be 22 mm (78%) greater behind the wheatgrass strips than in the open alfalfa field. This extra meltwater increased forage yield by 40% on the alfalfa field.

Relationships between snow accumulation and elevation were improved in western Canada where the influence of elevation on snow accumulation is most apparent at elevations above 600 m. Studies in the Monashee and Selkirk Mountains, British Columbia, show that the increase in accumulation with elevation is more dramatic where wind flows ascend mountains rather than where a descending flow occurs (Auld, 1995). Further north, in the southern Yukon Territory, the increase in snowfall with elevation is not apparent for elevation ranges from 600 to 1500 m (Pomeroy and Gray, 1995; Pomeroy *et al.*, 1999b), possibly due to more consistent subfreezing temperatures along the altitudinal gradient. A complication between elevation

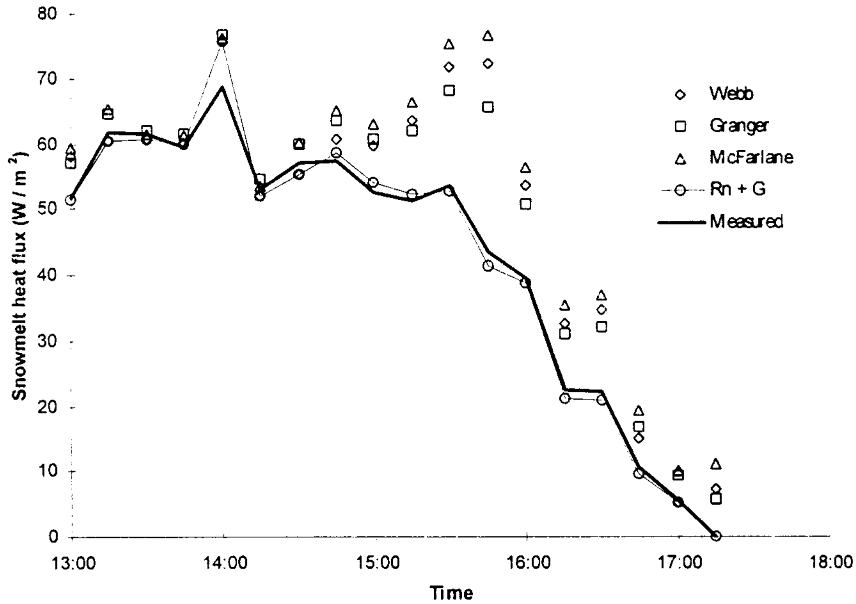

Figure 4. Comparison of snowmelt heat flux determined from measured inputs to the energy balance equation, and modelled using ground and radiation fluxes and estimated sensible and latent heat fluxes from the following schemes: Webb (1970), Male and Granger (1981) and McFarlane *et al.* (1992). The net radiation less ground heat flux is shown for further evaluation of the actual contribution from turbulent fluxes. 12 March 1996, near Saskatoon

and SWE relationships is noted by De Scally (1996), who quantified redistribution of snow by avalanches to lower elevations in southern British Columbia.

A recent technical advance is the use of weather radar to predict areal snowfall as an input to land surface and hydrological models (Fassnacht *et al.*, 1998). Weather radar has been evaluated and used in southern Ontario with encouraging results for providing a much better spatial representation of snowfall than is available with point measurements at weather stations. The authors were able to use weather radar snowfall as the precipitation input in calibrating a coupled land-surface–hydrological model, WATFLOOD/CLASS (Verseghy, 1991; Kouwen *et al.*, 1993). Calibrated streamflow simulations during snowmelt matched observations reasonably well in the Grand River, although the authors note that snow and frozen ground process improvements are necessary in the CLASS land surface scheme (Fassnacht *et al.*, 1998).

SNOWMELT HYDROLOGY

Recent research has emphasized the role of spatial heterogeneity in controlling surface energy fluxes and therefore melt, and the percolation of meltwater through layered snowpacks. This section examines these aspects of snowmelt hydrology carried out in Canada.

Surface energy and mass fluxes

Field-based research aided by improved turbulent flux measurements enabled an advancement of our understanding on the energy fluxes over both simple and complex surfaces. Pomeroy *et al.* (1998a) compared eddy correlation measurements of latent and sensible heat fluxes over a continuous, isothermal melting prairie snow cover with modelled exchanges from various popular formulations for these fluxes (Figure 4): the log–linear form used for stable conditions (Webb, 1970), a modification developed for snow cover by Male and Granger (1981), and a Richardson number formulation commonly used in atmospheric models (McFarlane *et al.*, 1992). All the turbulent transfer schemes overestimated the downward convective energy available for snowmelt, with root-mean-square errors of 10–13 W/m^2. These results suggest that existing formulations for convective transfer between atmosphere and snow cover should be examined further with respect to their performance under stable conditions. Of interest during the melt period was that the conductive heat flux into the frozen soil was small (<10 W/m^2), reflecting that previous strong temperature gradients between isothermal snow and frozen soil had diminished shortly after infiltration to unsaturated frozen soils (Pomeroy *et al*. 1999a). This finding is consistent with finite-difference scheme simulations from a coupled heat and mass transfer model for infiltration into frozen soils, which suggest that latent heat release upon freezing of infiltrating meltwater in unsaturated frozen soils will rapidly reduce temperature gradients and heat flux from snow to soil (Zhao *et al*., 1997).

Woo *et al*., (1999) examined the representativeness of point determinations of sensible or latent heat over snow-covered terrain and indicated that point values may not be extrapolated over large areas in the Arctic. Marsh (1999) reviewed various tower and aircraft eddy correlation methods for measuring spatially averaged fluxes and provided examples of applying these techniques to snow-free terrain. These techniques have yet to be applied to snow covered areas. To overcome the difficulties of measuring spatial variations in surface energy fluxes, it is often necessary to combine limited data sets with the use of models to fully represent the spatial variability. Several such approaches have been used.

Wind flow. Spatial variation in turbulent fluxes is at least partially dependent on variations in wind speed. Marsh (1999) summarized several methods for estimating wind fields over rolling terrain. They vary from physically based atmospheric models (e.g. Essery *et al*., 1999), to interpolation using extensive wind speed and direction observations, and interpolation using wind speed and direction observations in conjunction with empirical wind–topography relationships. These methods can be used for estimating variations in turbulent fluxes of sensible and latent heat over a melting snow cover, but no such attempt has yet been made. Such methods cannot be applied easily to situations that include alpine environments, patchy snow cover and conditions beneath a forest canopy. For example, the model used by Essery *et al*., (1999) for estimating wind speeds is applicable only for terrain with slopes that are less than 1 in 4, and obviously would not be appropriate to alpine environments.

Atmospheric boundary layer models. Atmospheric boundary layer models offer an alternative approach for considering the spatial variations of surface fluxes. Two-dimensional versions of such models have been used by Marsh *et al*. (1999) to model energy fluxes over snow, and three-dimensional models by Taylor *et al*. (1998) to consider thermally induced flows ('snow breezes') between a lake and a forest in the boreal environment. These models allow the determination of the turbulent exchanges of heat and mass over various land surfaces, hence permitting the modelling of meso- and local scale advection.

Local scale advection. The snow cover of open, windswept environments typically has highly variable snow depths, which quickly become patchy during melt (Figure 5). In addition, such environments often have patchy snow covers during late autumn for early winter, and in cases where the winters are long, cold and have low snowfall, patchy snow covers may persist throughout the winter (Shook and Gray, 1997b). Such patchy snow conditions are common in the Arctic tundra (Marsh and Pomeroy, 1996), grasslands and agricultural fields (Donald *et al*., 1995; Shook, 1995; Shook and Gray, 1997b), and beneath forest canopies (Donald *et al*., 1995; Faria, 1998). There are often large differences in albedo, surface roughness and surface temperature between the snow and snow-free patches. Although net radiation and sensible heat from regional scale advection provides the primary source of melt energy for melt in open environments, local advection of sensible heat from the snow-free areas to snow patches can significantly increase melt rates, as shown in the Prairie Snow Ablation Simulation (PSAS; Shook, 1995; Shook and Gray, 1997a). Local scale advection is further complicated by the gradually changing area and size distribution of the snow patches during the melt period. Mesoscale advection can be important in areas where surface heterogeneity occurs at such a

Figure 5. Air photograph showing patchy snow typical of the melt period of tundra terrain in northern Canada. This photograph was taken on 25 May 1993, approximately 50 km northwest of Inuvik, NWT

scale. Taylor *et al.* (1998) provided an example where thermal contrasts between snow-covered lakes and their surrounding boreal forest induced air flows that affected the low-level wind fields.

Marsh and Pomeroy (1996) and Neumann and Marsh (1998) used field measurements to show that the portion of the sensible heat advected from the bare areas to the snow patches declined with decreasing snow covered area. Similar results were obtained from a two-dimensional model (Marsh *et al.*, 1999). The number of field experiments remains limited, however, due to the difficulties of conducting field studies on local advection over snow. Shook (1995), for instance, found it difficult to measure changes in sensible heat flux along a downwind transect over a snow patch because the snow patches are irregular in shape and the fetch length changes rapidly with shifting wind directions. Consequently, research has focused on modelling (Shook, 1995; Shook and Gray, 1997a; Marsh *et al.*, 1999; Neumann, 1999). The model results confirmed that local advection of sensible heat increases melt, with the greatest effect immediately downwind of the snow-patch leading edge. Marsh *et al.* (1999) confirmed that the influence of advection decreased exponentially with distance from the leading edge to quickly reach a near steady state. However, estimates from field and modelling studies agreed only to within 25%. This distance-decay relationship is controlled primarily by wind speed, surface roughness and the difference in temperature between the snow-free and snow-covered areas.

Forest snowmelt

Snowmelt under forest canopies is strongly altered from the open environment case because the overlying canopy intercepts radiation and suppresses turbulent transfer. As a result, melt rates are lower under forests than in equivalent open areas, a factor used in Canadian forest water management for many decades. A complementary survey of forest snowmelt research is given by Buttle *et al.* (2000, this issue).

Pomeroy and Dion (1996) examined the role of canopy leaf area in extinguishing shortwave radiation and the relationship between this extinguished shortwave and subcanopy net radiation during snowmelt. They found that daily mean solar angle exerted a strong control on light extinction in pine canopies and therefore on subcanopy net radiation. In a medium density pine canopy in central Saskatchewan, subcanopy net radiation remained negative until late March and its magnitude was controlled by solar angle and insolation rather than air temperature. Metcalfe and Buttle (1995, 1998) examined the energy balance for snowmelt under boreal forest canopies in northern Manitoba, focusing on the role of canopy in affecting energy fluxes available for snowmelt. In their first study (1995) they found a liner relationship between the time to complete melt and canopy coverage and that melt was terminated at an open fen site, 12 days before completion at a dense black spruce site. Their second study (1998) refined this relationship to an exponential function between snowmelt rate and the non-canopy coverage (gap fraction). This statistical relationship was then used to map snowmelt over a heterogeneous area using canopy coverage from classified LANDSAT images.

Pomeroy and Granger (1997) measured the energetics of melt in stands of mature pine, mixed aspen–spruce, regenerating pine and clearcut in a Saskatchewan boreal forest. They found that subcanopy radiation dominated the snowmelt energy balance, with convective terms tending to cancel out in all environments. The largest net convective terms were measured under the mature pine canopy and accounted for 22% of total snowmelt energy during melt. For a total melt duration of 15 days, the convective term contributed to melt for the first 10 days in the pine canopy, first 5·5 days in the mixed-wood, first 2·5 days in the regenerating pine and for only the first few hours in the clearcut zone. A threefold increase in snowmelt rate upon clearcutting a mature boreal forest was thus inferred (Pomeroy *et al.*, 1997b).

Faria (1998) found that energy flux modification by dense canopies greatly delayed the depletion of snow cover in a boreal forest; the disappearance of snow advanced somewhat because of the covariance of melt energy and SWE, notably under mixed wood and medium density canopies. The covariance is due initially to decreasing interception of snow and decreasing melt energy (probably both radiation from trunks and sensible heat) with distance from tree trunks. As melt progresses the covariance is due to increasing SWE and decreasing melt energy (probably sensible heat) with distance from the melting edge of the snowpack (Faria, 1998).

Water routing and snow properties

Marsh (1991, 1999) noted several difficulties with the physical determination of water routing through snow, such as the possibility of non-Darcian flow in coarse grain saturated snow on steeper slopes and the effect of heterogeneity on the flow pattern. We are not yet able to quantify all the pertinent snow parameters that influence the flow through snow, but advancements have been made in recent years, as outlined in the following sections.

Flow variability and heterogeneity

A number of studies (see Marsh 1991, 1999) have shown that meltwater flux through snow is heterogeneous, with flow volumes being highly variable over small horizontal distances (e.g. Marsh and Pomeroy (1999) found that within a 0·25 m^2 area flow varied from zero to three times the mean value). A full appreciation of subsurface flow heterogeneity is still not attained, however. There are a variety of techniques used to quantify flow pathways, such as flow fingers and ice columns, which have been demonstrated to be highly variable in size and in spacing. McGurk and Marsh (1995), using a thick section technique to describe the mean flow finger spacing and diameter, showed that flow features are often not distributed uniformly, but tend to cluster in groupings. Although such clustering is typical, they have not been captured properly by appropriate statistical techniques.

SNOW CHEMISTRY

Snow chemistry as reviewed here includes research on the solid, liquid and gaseous phase chemical constituents of seasonal snow covers and their relationship with physical snow phenomena or processes that are relevant to hydrology. A new emphasis on the study of snow chemistry in Canada includes its important ecological interactions, particularly with respect to nutrient cycling (Jones, 1999). Jones reviewed the interactions between snow and climate, microorganisms, vegetation and animals. The influence of snow on nutrient cycling and its importance in the overall biological productivity of snow-covered ecosystems was illustrated by the study of the dynamics of N in biological, aqueous and gaseous forms (Jones, 1999).

The chemistry of accumulating snow cover is influenced by forest cover, as demonstrated by Hudson and Golding (1998) in British Columbia and Pomeroy *et al.* (1999c) in Saskatchewan and the Northwest Territories. Hudson and Golding (1998) found enriched concentrations of NO_3^-, SO_4^{-2}, Cl^- and Na^+ in snowpacks under subalpine canopies, compared with their concentrations in snowfall. Enrichment increases with increasing canopy density but the rates differ among different ions. Pomeroy *et al.* found enriched concentrations of SO_4^{-2} and Cl^- in snowpacks under boreal forest canopies, though little enrichment in NO_3^- concentrations. Increases in enrichment with coniferous leaf area was associated with concentration resulting from sublimation of the intercepted snow. In contrast to concentration, snowpack ion load relative to snowfall inputs was conserved by SO_4^{-2} and Cl^-. However, the loss of NO_3^- load was proportional to the sublimation loss of the intercepted snow, possibly as a result of volatilization of NO_3^- concomitant with destruction of snow crystals through sublimation (Pomeroy *et al.*, 1999b).

Snow chemistry can be modified by wind redistribution of snow. Landscape roughness and mesoscale wind exposure control snow chemical redistribution and loading, as was demonstrated by Pomeroy *et al.* (1995) in a transect across the tree-line near Inuvik, NWT. Pomeroy and Jones (1996) studied the effect of blowing snow on snow chemistry modifications in polar regions, yielding results that have significant implications for the interpretation of ice-core records. Through measurements and modelling, they showed that transport and sublimation of blowing snow particles can proportionately concentrate certain conservative ions (e.g. SO_4^{-2}, NH_4^+), enhance concentrations of scavenged ions (e.g. Cl^-, Na^+, Mg^{++}) and result in volatilization of NO_3^-. For conservative ions, a concentration increase resulting from blowing snow sublimation increases exponentially with wind speed. In polar regions, a threshold condition between snow ion retention with enhanced concentration and snow ion erosion and resuspension occurs at mean monthly 10-m wind speeds of 8 m/s.

Human activities can also influence the chemical composition of snow. Jones and Devarennes (1995) examined the chemical composition of artificial snow on a ski slope and its effects on plant germination. Snow density was greater (450 versus 250 kg m^{-3}) and concentrations of Ca^{2+} and HCO_3^- were much greater (355 versus 25 and 440 versus 0 meq m^{-3}, respectively) in artificial snow compared with natural mountain snow. Labadia and Buttle (1996) found NaCl concentrations in snow adjacent to major roads in southern Ontario to reach 9400 mg l^{-1} in late winter. The snowpack retained less than 50% of all applied salt during the winter.

Snow property research has been strongly linked to the study of snow chemistry in the last two decades. Advances in this area include that of Hoff *et al.* (1998), who developed a N-absorption technique to determine the specific surface area of snow; the technique may be useful in the studies of both snow properties and snow chemical processes. Snow cover properties affect the transmission of gas through snow, as demonstrated by van Bochove *et al.* (1995, 1996), who measured profiles of CO_2 and N_2O in Quebec agricultural and forest snowpacks. Winter N loss from the soil through snow was 5·8 kg/ha N–N_2O for open snow covers but negligible from snow under forests. Van Bochove *et al.* found a three- to fivefold increase in CO_2 and N_2O gas flux during snowmelt when compared with the accumulation period. Jones *et al.* (1999a) used a gas diffusion system to measure the diffusion coefficient for gas in snow and the resistivity of snow to gas transfer (if less than 1·0, diffusion is restricted). They found that resistivity declined from 1·0 to 0·6 as the winter season progressed.

Marsh and Pomeroy (1999) conducted research in the Arctic on the impact of flow fingering and flow variability on snow chemistry, and found that chemicals are eluted first through finger flow and secondly through matrix flow. As a result, enhanced concentrations of ions during early melt (fractionation) are partly due to the combination of heterogeneous flow through the snowpack and initial solute location in the snow–water mixture. This finding partly explains observations of ion fractionation and preferential elution made in the previous decade and compiled by Jones and Orville-Thomas (1987). The effect of mid-winter rainfall

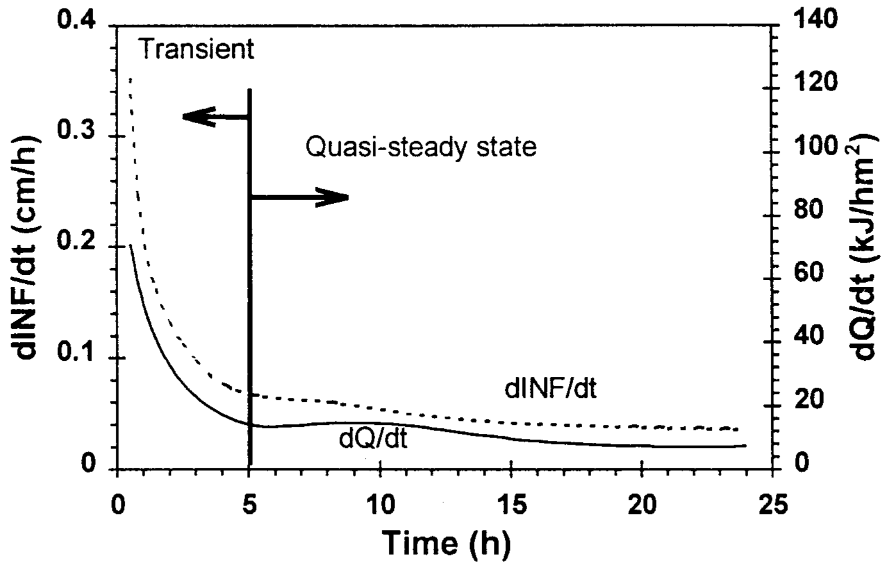

Figure 6. Variations in infiltration rate, dINF/dt, and surface heat flux rate, dQ/dt, with time during snowmelt infiltration into a frozen silty clay soil (Zhao and Gray, 1997a)

on snowmelt chemistry was investigated by Mclean *et al.* (1995), who found that because of fractionation, small rainfall events removed over 60% of the ionic load but only 7% of SWE. Hudson and Golding (1998) detailed further observations on the rate of depletion of ions from snowpacks during melt in forested and clearcut sites in British Columbia and demonstrated the influence of forest cover on rates of preferential elution.

Although most chemistry studies have examined small-scale processes or plots, the snow chemical load is involved in the major annual runoff events in many Canadian environments. Snowmelt runoff chemistry on hillslopes can also relocate the solute. Hayashi *et al.* (1998) found that 4–5 kg/year of Cl was transported in snowmelt water from an upland to a slough (small prairie wetland) in a Saskatchewan pastureland to form an important part of the annual Cl cycle.

FROST HYDROLOGY

Frost hydrology studies the distribution, storage and movement of water under the direct and indirect influence of seasonally or perennially frozen soils. This survey emphasizes the research conducted in areas with permafrost, with the exception of infiltration studies carried out in the Prairies and boreal forest where seasonal frost occurs, with great importance for the generation of spring runoff.

Infiltration into frozen soils

The Division of Hydrology, University of Saskatchewan expanded its research on frozen soil infiltration through theoretical studies and quantification of the thermodynamic processes, and by extending the Prairie field studies carried out in the 1980s (Granger *et al.*, 1984) to boreal and subarctic regions (Zhao and Gray, 1999). Zhao *et al.* (1997) derived a numerical analysis of phase change and coupled heat and mass transfer into homogeneous, unsaturated frozen soils. The results suggest that frozen soil infiltration has two primary regimes as meltwater is applied to a frozen soil: an early transient regime with higher soil heat fluxes and infiltration rates and a later, more subdued, quasi-steady-state regime (Figure 6). Once the quasi-steady-state is reached the energy available to increase soil temperature at depth is supplied by latent heat released through freezing of infiltrating water in the upper soil layers. Zhao and Gray (1997a) used this numerical simulation to derive a parametric expression that estimates cumulative infiltration after the attainment of quasi-steady-state flow, using surface saturation, initial soil moisture content and soil temperature, and the product of hydraulic conductivity and time as the dependent variables. The most important variables controlling infiltration are infiltration opportunity time (a function of snow water equivalent) and initial soil moisture content. Soil texture alone had little effect on frozen soil infiltration (Zhao and Gray, 1997b). Tests of the parametric equations suggested reasonable correspondence with field measurements in the Prairie and boreal forest regions of Canada, the difference in infiltration between sites is probably due to the lack of ponded water in melting boreal forest snowpacks (Zhao and Gray, 1999).

In terms of field application, Harms and Chanasyk (1998) found it difficult to apply the empirical relationship between infiltration, snow water equivalent and pre-melt volumetric moisture content presented by Granger *et al*. (1984) to predict infiltration using maximum snow water equivalent and autumn soil moisture as the inputs. They suggested that mid-winter melts alter near-surface soil moisture and snow accumulation and hence the autumn values should not be used in this simulation. Recent more physically-based model developments should provide a means to avoid this problem by using pre-melt soil moisture conditions (Zhao and Gray, 1999).

Groundwater and icing

In the continuous permafrost region of Canada, the impervious permafrost usually separates groundwater occurrence into suprapermafrost and subpermafrost zones. The exfiltration of groundwater from below the permafrost is rare. One exception is the perennial springs found in western Axel Heiberg Island of the High Arctic (Pollard *et al*., 1998). These springs are found in two clusters, discharging warm (3·5–6·7°C) and highly mineralized water that may be relict in origin. More prevalent is the exfiltration of groundwater that resides above the permafrost. Woo and Xia (1996) reported seepage in coarse gravelly terrain in the High Arctic, which supports local saturated zones, patchy wetlands, ponds and runoff in small streams. The contribution of groundwater discharge to surface runoff in the periglacial environment is now included in text-book materials (French, 1996).

Whereas the suprapermafrost groundwater source may be depleted by the end of summer, intra- and subpermafrost sources are less prone to seasonal variations so that discharge is maintained during the winter in the discontinuous permafrost areas, where taliks (perennial unfrozen zones) provide conduits for the groundwater to move to the surface. Icings (aufeis in German) are formed when successive seepage flows of water are frozen by the cold winter air, producing sheet-like masses of layered ice. Recent studies on the growth and decay of icings were carried out in subarctic Yukon and southwestern Northwest Territories. Clark and Lauriol (1997) examined the calcite precipitates from icings and deduced that the waters that contribute to baseflow and icing formation are recharged through an unsaturated zone underlain by talik rather than by a permafrost substrate. Hu and Pollard (1997a) modelled the growth of icings from a water balance perspective. The first stage of icing formation is the freeze-up, when a river ice cover is established and is anchored to the river banks. The second stage is ice growth involving the thickening of the river water, the inclusion of snow in the ice and the freezing of water discharged to the icing section. The final stage is growth by the freezing of overflow and the incorporation of any snow that is deposited on the ice cover. To investigate the detailed variations in icing topography resulting from overflow, Hu and Pollard (1997b) set up a field experiment near Montreal to measure the growth of ice by discharging water to a 10 m × 1 m plot. They found that simultaneous growth can occur only over a limited distance, and growth rate at any point is autocorrelated only with ice accumulation during the previous period. These results suggest that there is constant shifting in the position of icing growth. Multiple regression indicates that the spreading of the icing layer is governed by discharge, water and air temperature, wind speed and slope of the pre-existing icing surface.

The ablation of icing contributes to streamflow during the snowmelt period. Reedyk *et al*. (1995) found that 5% of the total flow of a small sub-Arctic stream during the snowmelt high flow season is supplied by icing melt. Continued ablation of icing after the disappearance of snow extends the melt season for streamflow generation. In the sub-Arctic, the ablation rate is higher than in the colder Arctic, where slower icing melt sustains streamflow for a longer duration. For the Firth River in the discontinuous permafrost belt of north-western Yukon Territory, Clark and Lauriol (1997) estimated that icing represents over 30% of annual groundwater discharge, and groundwater provides about half of the river flow. Such a high percentage is comparable to values reported for non-permafrost carbonate rock terrain.

Wetlands

The High Arctic has many wetlands that occur in patches. Their presence is due to focused water supply and a shallow frost table (Woo and Young, 1998). The latter condition may be attributed to the feedback between ground-ice formation and frequent saturation: saturation produces abundant ice in the active layer in the winter but summer ground thaw is retarded by the large latent heat consumption for ice melt, thus limiting the frost depth (Woo and Xia, 1996). Ice-rich frost inhibits deep percolation of snow meltwater and rainfall, thus maintaining frequent saturation and the presence of wetlands.

In the spring, the largely frozen wetland has little storage capacity to absorb water input, with the consequence that the wetland serves as a pathway for the delivery of meltwater and has limited attenuation effect for high flows (Glenn and Woo, 1997). Summer evaporation reduces water storage and may increase the effectiveness to attenuate storm runoff, although this depends on the antecedent moisture, the amount of rainfall and the replenishment of water supplies through inflow and ground-ice melt.

Slope and catchment hydrology

Young *et al.* (1997) studied four slopes of different orientation in the Arctic tundra (80°N) and found their precipitation distribution, snowmelt, frost table configuration, water table fluctuations and slope flow rates to differ. The contrasts in radiation, air and ground temperatures were exaggerated during a warm, dry summer, causing larger variations in thermal and hydrological responses among slopes compared with a cloudy summer. Intraslope variations in topography and vegetation led to localized diversities. Kokelj and Lewkowicz (1998) noted that the presence of active-layer detachment slides increases snow accumulation within the scar zone and the runoff response to rainfall is also greater at the landslide sites than the adjacent vegetated areas.

Hydrological contrasts between two slopes are more pronounced in the sub-Arctic than in the Arctic. Carey and Woo (1998) compared two forested sub-Arctic slopes (61·5°N): a north-facing slope underlain by permafrost and covered by spruce open woodland and a south-facing slope with a dense aspen cover and only seasonal frost. During the snowmelt season, melting on the south slope was advanced by weeks and the snow either sublimated or infiltrated the seasonal frost but yielded no runoff and possibly adding little to raise the local water table. The north slope has icy frost beneath a top layer of peat, the former is relatively impervious whereas the latter is rapidly saturated to produce rapid lateral flows within the peat, along soil pipes or overland.

Further south in central Alberta (53·5°N), Harms and Chanasyk (1998) also reported thermal and hydrological differences between north- and south-facing slopes covered by forage as parts of a reclaimed mine. The south slopes experienced early snowmelt, received larger gain in their soil moisture during winter and yielded less runoff than the north slopes.

All these studies demonstrate that slope orientation gives rise to local differences in slope hydrology and though the absolute magnitudes vary from year to year, there is a recurrent spatial pattern of snow accumulation, snowmelt, ground frost, subsurface moisture storage and slope runoff. The persistence of microclimatic and hydrological contrasts is a major consideration in the development of vegetation, ground frost and even soils. Another implication, particularly notable for the sub-Arctic forested environment, is that only some slopes contribute runoff to streamflow (Carey and Woo, 1999). Thus, only portions of the drainage basin are source areas for flow generation. In terms of upscaling, the slope studies demonstrate that variability on a local scale can be pronounced and the representativeness of point-specific observations should be assessed before such data are extrapolated to an area (Young *et al.*, 1997).

Slope runoff mechanism is examined in detail by Quinton and Marsh (1998a, 1999) at a tundra site (68·7°N) that is covered by earth hummocks, a form of patterned ground widely distributed in permafrost terrain. Whereas the peat-filled interhummock zones concentrate and facilitate flow because of the high hydraulic conductivity of the organics, the hummocks obstruct flow from following a direct path to the streambank. Flow is attenuated by storage in the interhummock area, causing a rise of the water table in the peat. The retardation of flows in this tundra environment during the snowmelt period is not only due to delays on hillslopes (Quinton and Marsh, 1998b). The delivery of meltwater lags behind snowmelt because vertical flow through the snowpack is initially very slow. When the snow is thoroughly wetted, rapid percolation is accompanied by fast slope runoff, but lateral flow slackens as the active layer thaws. Slope runoff may accumulate at the bottom of deep snow that fills the stream channel. Streamflow velocity increases only after the snow blockage is removed by thermal and mechanical processes (Woo and Young, 1997; Quinton and Marsh, 1998b).

Janowicz *et al.* (1997) examined snowmelt runoff generation from a sub-Arctic mountainous catchment (Wolf Creek, Yukon Territory) and found that the spring freshet was not related directly to average basin snow water equivalent but was strongly affected by the synchronicity of snowmelt in various elevation–vegetation bands and by autumn water budget. The highest peak streamflow events were associated with concurrent, rapid snowmelt across the elevation bands; whereas the magnitude of spring discharge was related to autumn soil moisture. The authors calculated that variation in frozen soil infiltration capacity could account for the variation observed in spring discharge volumes in years with similar maximum snow accumulations. The combination of differences in snowmelt synchronicity and soil moisture resulted in a 10-fold difference in peak spring discharge over 3 years, despite similar winter snowpacks.

Snow is the major contributor to streamflow and spring melt is the period with the highest runoff of the year. Large year-to-year variations in winter snowfall leads to a corresponding variation in streamflow and in the recharge of basin storage (Woo and Young, 1997; Carey and Woo, 1998). Under a polar oasis climate, evaporation is higher than the polar desert condition and evaporation is enhanced accordingly. Increased thaw of the active layer in the summer increases basin storage capacity and much of the rainfall goes to soil moisture replenishment instead of streamflow generation.

RESEARCH DIRECTIONS

Process studies

In spite of the advances made in the last decades, many physical processes related to snow and frost are still not well understood or adequately quantified. Improved measurement techniques and theoretical studies have also identified serious misconceptions in the descriptions of certain processes that were felt to be well-understood. Intensive field and laboratory studies of cold region hydrological processes should continue, based on rigorous application of physical principles, and particularly paying increased attention to the interaction of energy and mass flows between atmosphere, snow and soils and to conservation of mass and energy in dynamic systems subject to phase change. Snow and frost process descriptions need to be improved for heterogeneous terrain, particularly mountains and forests.

As examples, more work is needed to predict accurately the flow of water and heat in snow and soil, including the effects of flow fingers, the formation of ice layers and their influence on heat and water transfer. Fundamental experimental work on infiltration into frozen soils should also consider the case of heterogeneous infiltration. The results of such experiments will surely differ from the ideal homogeneous cases that models can presently consider. The coupling of snowmelt and frozen ground energetics and exchange with atmospheric flows is at a primitive stage and could yield new information on advective contributions to snowmelt energy.

Boundary considerations

The existence of boundaries, both above and below ground, affects snow and frost-related hydrological processes and fluxes. *Fixed boundaries* such as tree-line, ground surface or lake border can be found in all climatic zones and the cold regions are no exception. *Shifting boundaries* change with the season; examples include the edge of snow cover, depth of snow and the frost table (top of the permanently frozen zone). *Vanishing boundaries* are an extreme variety where particular features enclosed by the boundaries disappear. Examples include the complete melting of the snow cover or the thawing of the seasonal frost. *Fluid boundaries* are dynamic and occur at the edge of a mass of air or water that moves across an area (e.g. the front of a body of cold water sinking into a lake, or the leading edge of warm air advecting over a snow patch). The study of atmospheric fluid boundaries is encompassed by the subfield of boundary layer meteorology. The occurrence of boundaries has significant hydrological implications because they:

1. demarcate a change in the characteristics of the medium in which hydrological processes occur, e.g. hydraulic conductivity is different between the frozen and the thawed soils within and outside frozen zone;
2. are crossed by fluxes of mass and energy that affect the rates of hydrological processes, e.g. increased trapping of blowing snow at the forest margin or accelerated snowmelt at the edge of a snow patch (Marsh and Pomeroy, 1996; Pomeroy *et al.*, 1997b; Neumann and Marsh, 1998);
3. shift and therefore affect hydrological processes and state variables, e.g. water table in the active layer falls as the frost table deepens, but in winter, the descent of a freezing front may force the groundwater to rise above ground to be frozen as icing;
4. require preservation of the continuity of mass and energy at multiple scales — this concept is not always applied rigorously in hydrology process studies and models.

These effects are reasonably well-understood in certain systems but can become exceeding complex where considerations of mass–energy continuity, mass–energy flux and phase change are required. Hence in many cold hydrological systems they remain to be analysed theoretically and quantified through field measurement. The processes leading to the shifting or disappearance of boundaries have to be examined and modelled accurately. In addition, distributed models often superimpose *artificial boundaries* (e.g. grid-lines) on to the hydrological system. These boundaries have significant complicating effects on hydrological computations.

Modelling cold region processes

There is still a need to develop hydrological models that explicitly reflect the snow and frozen soil processes important to northern Canada. Physically based modelling of frozen soil processes is needed to represent correctly and in adequate detail the dynamics of the seasonal freeze–thaw zone and their effects surface–subsurface hydrology, especially snowmelt (Pomeroy *et al.*, 1998a). Increasing demand by land surface schemes and large collaborative projects (e.g. GEWEX, see Stewart *et al.*, 1998) calls for hydrological models for cold regions on a macroscale. Available largescale models incorporate equations adapted to the temperate latitudes, supplemented by empiricism based on limited data and produce outputs that are seldom verifiable. Models developed in southern Canada have been applied to areas with discontinuous permafrost (Kite *et al.*, 1994; Pietroniro *et al.*, 1996). Semi-distributed versions of the SLURP (simple lumped reservoir parametric) model and SPL7, the hydrological model of WATFLOOD (a hydrological database management system) have been used to simulate runoff of the Mackenzie basin and several wetland-dominated tributary catchments of the Liard River. These models calibrate their parameters by optimization but are limited by the sparse data network for the northern region. Pietroniro *et al.*, (1996) identified possible improvements by replacing temperature index with radiation input for snowmelt computation, by introducing the effects of antecedent moisture and frozen ground conditions to modify hydraulic conductivity and interflow rate coefficients, and by taking into consideration the presence of icing and peat. Pomeroy *et al.* (1998c) added frozen oil infiltration and blowing snow redistribution (PBSM) to SLURP and found dramatic improvements when used to simulate spring runoff in a Prairie basin and a reduced need to calibrate model parameters when producing runoff hydrographs. Continued improvement of cold region process representation in hydrological models will enhance model performance significantly.

Spatial variability and scale issues

In terms of energy and mass fluxes, significant instrument advances (including eddy correlation and aircraft flux measurements) have provided estimation of their spatial variability in surface fluxes, but these techniques have not been exploited fully for snow covered areas. Furthermore, although there are boundary layer models that consider local and mesoscale advection under snow-covered conditions, these and other techniques such as rough terrain wind-flow models, have not been fully utilized to assess the spatial variations in energy fluxes. Many of these techniques should be explored to obtain a better understanding of energy and mass fluxes over the snow, and their applications should be extended to complex, steep terrain.

The scale and scaling issues are of increasing importance in view of parameterization and data input to distributed models and land surface schemes in climatological models. A critical question for snow and frost studies is the appropriate scale at which processes may be represented. Fractal mathematics has provided an efficient numerical tool for describing snow processes at variable scales, but can be used further in scaling research. Canada lags behind others in theoretical and applied investigations on scale issues in frost-related hydrology (cf. Kirnbauer *et al.*, 1994; Blöschl, 1999). As an example, there is little understanding about covariances between the driving parameters for snowmelt infiltration to frozen soils. Woo (1998) noted that different types of snow information may be needed for hydrological investigations carried out at different scales; but very often, data collected at a local scale are used for large-scale hydrological studies without undergoing proper up-scaling. This can produce erroneous representations of the snow conditions for grid-cells commonly used for regional climate modelling (Yang and Woo, 1999). Snow processes should be up-scaled with reference to their interactions over time and space and to the applicability of spatial statistical distributions as scale is increased.

Interactions with geochemistry and ecology

Heat and moisture, as affected by snow, are important to and perhaps define the distinctiveness of ecological and chemical processes in Canada. An international Snow Ecology Working Group of the International Commission on Snow and Ice hosted by Canada has outlined the basis for consideration of snow as an ecosystem and of snow ecology as a field of study analogous to limnology (Jones *et al.*, 1994), but broadly based research on snow ecology has only begun (Jones, 1999). A working group, led by Canadians, produced a synthesis of the present state of the prospects for the field of snow ecology (Jones *et al.*, in press). In cold regions, more work should be done to understand the role played by the snow cover and frozen soils in controlling the geochemistry of streams, soils and water bodies. Plant productivity and chemical reactions are notably influenced by seasonal frost and permafrost. The position of the suprapermafrost water table in northern wetlands, for example, is found to be controlling factor in the production of greenhouse gases, such as methane and carbon dioxide (Waddington *et al.*, 1998). Although biological investigations in cold regions often need inputs from physical hydrology, vegetation contributes significant feedbacks to the hydrology of snow and frost, the former includes the effects on snow accumulation, redistribution, albedo and melt, the latter through aerodynamic roughness, evapotranspiration and changes in thermal properties of the near-surface layer owing to plant growth and decay. Canada will host the Snow–Vegetation Interactions Working Group of the International Commission on Snow and Ice starting in 1999, to address the issues related to snow and its relationship with plants. Through this type of effort and interactions with various disciplines, it is hoped that snow and frost hydrology will become integrating foci of a comprehensive cold regions science.

CONCLUSIONS

This review of Canadian research in snow, frozen soils and permafrost hydrology has shown advancement on most of the issues that were identified at the beginning of the review period (1995). This is to be expected because the issues were focused. Advances have been particularly strong in the areas of snow and frozen soil processes and runoff generation from permafrost areas. A new perspective on snow studies has resulted from the development of the subfield of snow ecology, with particular strengths in Canada. Unfortunately, despite development of process algorithms, advances in comprehensive mathematical models to incorporate the impacts of the snow and cold region hydrological processes on hydrological and atmospheric systems have not been abundant in this period. Much remains to be done, as discussed in the extensive section on research directions. New research in the areas of cold region hydrological processes, influence of boundaries, process-based models, scaling and links with geochemistry and ecology is called for. The Canadian snow, frozen ground and permafrost hydrology community is changing with rapid institutional evolution in governments and universities. It is hoped that the next generation of research will pay detailed attention to mass and energy continuity, thermodynamics of phase change, rigorous application of physical and chemical principles and the role of snow in the ecosystem in advancing this quintessentially Canadian form of hydrology.

REFERENCES

Albert MR, Prowse TD, Taylor S. 1998. Preface: CGU Hydrology Section/Eastern Snow Conference *Hydrological Processes* **12**(10–11): 1525.
Auld H. 1995. Dependence of ground snow loads on elevation in western Canada. *Proceedings of the Western Snow Conference* **63**: 143–146.
Beltaos S. 2000. Advances in river ice hydrology. *Hydrological Processes* **14**: 1613–1625.
Blöschl, G. 1999. Scaling issues in snow hydrology. *Hydrological Processes* **13**: 2149–2175.
Brown RJE (ed.). 1969. *Proceedings of the Third Canadian Conference on Permafrost*, *14 and 15 January*, *1969*. Technical Memorandum No. 96, National Research Council of Canada: Ottawa, Ontario; 187 pp.
Buttle JM, Creed IR, Pomeroy JW. 2000. Advances in forest hydrology 1995–1998. *Hydrological Processes* **14**: 1551–1578.
Carey S, Woo MK. 1998. Snowmelt hydrology of two subarctic slopes, southern Yukon, Canada. *Nordic Hydrology* **29**: 331–346.
Carey S, Woo MK. 1999. Hydrology of two slopes in subarctic Yukon, Canada. *Hydrological Processes* **13**: 2549–2562.
Clark ID, Lauriol B. 1997. Aufeis of the Firth River basin, Northern Yukon, Canada: insights into permafrost hydrogeology and karst. *Arctic and Alpine Res*. **29**: 240–252.
CNC-IHD. 1970. *Snow Hydrology*, *Proceedings of Workshop Seminar*. Secretariat, Canadian National Committee for the International Hydrological Decade: Queen's Printer for Canada: Ottawa, Ontario; 82 pp.
De Scally FA. 1996. Avalanche snow melting and summer streamflow differences between high-elevation basins, Cascade Mountains, British Columbia, Canada. *Arctic and Alpine Research* **28**: 25–34.
Dery SJ, Taylor PA. 1996. Some aspects of the interaction of blowing snow with the atmospheric boundary layer. *Hydrological Processes* **10**: 1345–1358.
Dery SJ, Taylor PA, Xiao J. 1998. The thermodynamic effects of sublimating, blowing snow in the atmospheric boundary layer. *Boundary-layer Meterology* **89**: 251–283.
Donald JR, Soulis ED, Kouwen N, Pietroniro A. 1995. A land-cover based snow cover representation for distributed hydrological models. *Water Resources Research* **31**: 995–1009.
Essery R, Li L, Pomeroy J. 1999. A distributed model of blowing snow over complex terrain. *Hydrological Processes* **13**: 2423–2438.
Faria DA. 1998. *Distributed energetics of boreal forest snowmelt*. MSc thesis, Division of Hydrology, University of Saskatchewan: Saskatoon; 140 pp.
Fassnacht SR, Soulis ED, Snelgrove KR, Kouwen N. 1998. Application of weather radar to model the snow hydrology of southern Ontario. *Proceedings of the Eastern Snow Conference* **55**: 115–123.
French HM (ed.). 1982. *The Roger J.E. Brown Memorial Volume*, *Proceedings of the Fourth Canadian Permafrost Conference*. National Research Council of Canada: Ottawa; 594 pp.
French HM. 1996. *The Periglacial Environment*, 2nd edn. Longman: Harlow; 341 pp.
Glenn MS, Woo MK. 1997. Spring and summer hydrology of a valley-bottom wetland, Ellesmere Island, Northwest Territories, Canada. *Wetlands* **17**: 321–329.
Granger RJ, Gray DM, Dyck GE. 1984. Snowmelt infiltration to frozen Prairie soils. *Canadian Journal of Earth Science* **21** 669–677.
Gray DM, Male DH. (eds.). 1981. *Handbook of Snow: Principles*, *Processes*, *Management and Use*. Pergamon Press: Toronto.
Gray DM (ed.). 1970. *Handbook on the Principles of Hydrology*. Water Information Center: Syosset, NY.
Harding RJ, Pomeroy JW. 1996. The energy balance of the winter boreal landscape. *Journal of Climate* **9**: 2778–2787.
Hardy JM, Albert M, Marsh P. 1999. Preface: International Conference on Snow Hydrology, the Integration of Physical, Chemical and Biological Systems. *Hydrological Processes* **13**: 1719.

Harms TE, Chanasyk DS. 1998. Variability of snowmelt runoff and oil moisture recharge. *Nordic Hydrology* **29**: 179–198.
Hayashi M, van der Kamp G, Rudolph DL. 1998. Water and solute transfer between a prairie wetland and adjacent uplands, 2. Chloride cycle. *Journal of Hydrology* **207**: 56–67.
Hedstrom NR. 1998. *Development and evaluation of a cold regions snow interception model*. MSc thesis, Division of Hydrology, University of Saskatchewan: Saskatoon; 101 pp.
Hedstrom NR, Pomeroy JW. 1998. Measurements and modelling of snow interception in the boreal forest. *Hydrological Processes* **12**: 1611–1625.
Hoff JT, Gregor D, Mackay D, Wania F, Jia CQ. 1998. Measurement of the specific surface area of snow with the nitrogen adsorption technique. *Environmental Science and Technology* **32**: 58–62.
Hu XG, Pollard WH. 1997a. Ground icing formation: experimental and statistical analyses of the overflow process. *Permafrost and Periglacial Processes* **8**: 217–235.
Hu XG, Pollard WH. 1997b. The hydrologic analysis and modelling of river icing growth, North Fork Pass, Yukon Territory, Canada. *Permafrost and Periglacial Processes* **8**: 279–294.
Hudson RO, Golding DL. 1998. Snowpack chemistry during snow accumulation and melt in mature subalpine forest and regenerating clear-cut in the southern interior of BC. *Nordic Hydrology* **29**: 221–244.
Janowicz JR, Gray DM, Pomeroy JW. 1997. Snowmelt and runoff in a subarctic mountain basin. In *Proceedings of the Hydro-ecology Workshop on the Arctic Environmental Strategy*, Milburn D (ed.). Symposium No. 16, National Hydrology Research Institute: Saskatoon; 303–320.
Jones HG. 1999. The ecology of snow-covered systems: a brief overview of nutrient cycling and life in the cold. *Hydrological Processes* **13**: 2135–2147.
Jones HG, Devarennes G. 1995. The chemistry of artificial snow and its influence on the germination of mountain flora. In *Biogeochemistry of Seasonally Snow-covered Catchments*. IAHS Publication 228, International Association of Hydrological Sciences: Wallingford; 355–360.
Jones HG, Orville-Thomas WJ (eds). 1987. *Seasonal snowcovers: Physics, Chemistry, Hydrology*. NATO ASI Series C211, D. Reidel Publishing: Dordrecht; 746 pp.
Jones HG, Pomeroy JW, Walker DA, Wharton RA. 1994. Interdisciplinary group investigates snow ecology. *EOS, Transactions of the American Geophysical Union* **75**(14): 162–163.
Jones HG, van Bochove E, Bertrand N. 1999a. The transmission of soil gases through seasonal snowcover: an experiment to determine the diffusivity of N_2O in snow *in situ*. In *Interactions between the Cryosphere, Climate and Greenhouse gases*, Tranter M (ed.). IAHS Publication 256, International Association of Hydrological Sciences: Wallingford; 237–244.
Jones HG, Pomeroy JW, Walker DA, Hoham R (eds). In press. *The Ecology of Snow*. Cambridge University Press: Cambridge.
Kirnbauer R, Blöschl G, Gutknecht D. 1994. Entering the era of distributed snow models. *Nordic Hydrology* **25**: 1–24.
Kite GW, Dalton A, Dion K. 1994. Simulation of streamflow in a macroscale watershed using general circulation model data. *Water Resources Research* **30**: 1547–1559.
Kokelj SV, Lewkowicz AG. 1998. Long-term influence of active-layer detachment sliding on permafrost slope hydrology, Hot Weather Creek, Ellesmere Island, Canada. *Proceedings, Seventh International Conference on Permafrost*, Yellowknife, NWT, Centre d'ětudes nordiques, Laval University; 583–589.
Kouwen N, Soulis ED, Pietroniro A, Donald JR, Harrington RA. 1993. Grouped response units for distributed hydrologic modelling. *Journal of Water Resources Planning and Management* **119**(3). 289–305.
Kry PR (ed.). 1987. *Snow Property Measurement Workshop*. Technical Memorandum No. 140, National Research Council of Canada: Ottawa; 524 pp.
Labadia CF, Buttle JM. 1996. Road salt accumulation in highway snow banks and transport through the unsaturated zone of the Oak Ridges Moraine, southern Ontario. *Hydrological Processes* **10**: 1575–1589.
Lapen DR, Martz ML. 1996. An investigation of the spatial association between snow depth and topography in a prairie agricultural landscape using digital terrain analysis. *Journal of Hydrology* **184**: 277–298.
Li L, Pomeroy JW. 1997a. Estimates of threshold wind speeds for snow transport using meteorological data. *Journal of Applied Meterology* **36**: 205–213.
Li L, Pomeroy JW. 1997b. Probability of Blowing snow occurrence by wind. *Journal of Geophysical Research* **102**(D18): 21 955–21 964.
Male DH, Granger RJ. 1981. Snow surface energy exchange. *Water Resources Research* **17**: 609–627.
Marsh P. 1991. Water flux in melting snow covers. In *Advances in Porous Media*, Corapcioglu MY (ed.). Elsevier: Amsterdam; Vol. 1, 61–124.
Marsh P. 1999. Snowcover formation and melt: recent advances and future prospects. *Hydrological Processes* **13**: 2117–2134.
Marsh P, Pomeroy JW. 1996. Meltwater fluxes at an arctic forest-tundra site. *Hydrological Processes* **10**: 1383–1400.
Marsh P, Pomeroy JW. 1999. Spatial and temporal variations in snowmelt runoff chemistry, Northwest Territories, Canada. *Water Resources Research* **35**: 1559–1567.
Marsh P, Neumann N, Pomeroy J, Essery R. 1999. Comparison of model and field estimates of local scale advection of sensible heat over a patchy Arctic snow cover. In *Interactions between the Cryosphere, Climate and Greenhouse Gases*, Tranter M (ed.). IAHS Publication 256, International Association of Hydrological Sciences: Wallingford; 103–110.
McConkey BG, Ulrich DJ, Dyck FB. 1997. Snow management and deep tillage for increasing crop yields on a rolling landscape. *Canadian Journal of Soil Science* **77**: 479–486.
McFarlane NA, Boer GJ, Blanchet J-P, Lazare M. 1992. The Canadian Climate Centre second generation circulation model and its equilibrium climate. *Journal of Climate* **5**: 1013–1044.
McGurk BJ, Marsh P. 1995. Flow finger continuity in serial thick sections in a melting Sierran snowpack. *Biogeochemistry of Seasonally Snow-covered Catchments*. IAHS Publication 228, International Association of Hydrological Sciences: Wallingford; 81–88.
Mclean, RA, English MC, Schiff SL. 1995. Hydrological and hydrochemical response of a small Canadian Shield catchment to late winter rain-on-snow events. *Hydrological Processes* **9**: 845–863.

McNay RS, Petersen LD, Nyberg JB. 1988. The influence of forest stand characteristics on snow interception in the coastal forests of British Columbia. *Canadian Journal of Forestry Research* **18**: 566–573.

Metcalfe RA, Buttle JM. 1995. Controls of canopy structure on snowmelt rates in the boreal forest. *Proceedings of the Eastern Snow Conference* **52**: 249–257.

Metcalfe RA, Buttle JM. 1998. A statistical model of spatially distributed snowmelt rates in a boreal forest basin. *Hydrological Processes* **12**: 1701–1722.

Munro DS. 2000. Progress in glacier hydrology: a Canadian perspective. *Hydrological Processes* **14**: 1627–1640.

Neumann NN. 1999. *Local advection of sensible heat in the Arctic snowmelt landscape*. Unpublished MSc thesis. Department of Geography, University of Saskatchewan: Saskatoon; 120 pp.

Neumann N, Marsh P. 1998. Local advection in the snowmelt landscape of arctic tundra. *Hydrological Processes* **12**: 1547–1560.

Northern Research Basins. 1997. *Proceedings Eleventh International Symposium and Workshop*, Volumes 1 and 2, Prudhoe Bay, Fairbanks, Alaska, August; 249 pp. plus 394 pp.

Pietroniro A, Leconte R. 2000. A review of Canadian remote sensing applications in hydrology, 1995–1999. *Hydrological Processes* **14**: 1641–1666.

Pietroniro A, Prowse TD, Hamlin L, Kouwen N, Soulis R. 1996. Application of a grouped response unit hydrological model to a northern wetland region. *Hydrological Processes* **10**: 1245–1261.

Pollard WH, Omelon C, Andersen D, McKay C. 1998. Geomorphic and hydrologic characteristics of perennial springs on Axel Heiberg Island, Canadian High Arctic. *Proceedings*, *Seventh International Conference on Permafrost*, Yellowknife, NWT, Centre d'ëtudes nordiques, Laval University; 909–914.

Pomeroy JW, Dion K. 1996. winter radiation extinction and reflection in a boreal pine canopy: measurements and modelling. *Hydrological Processes* **10**: 1591–1608.

Pomeroy JW, Goodison BE. 1997. Winter and snow. In *The Surface Climates of Canada*, Bailey WG, Oke TR, Rouse WR. McGill-Queen's University Press: Montreal, 68–100.

Pomeroy JW, Granger RJ. 1997. Sustainability of the western Canadian boreal forest under changing hydrological conditions — I — snow accumulation and ablation. In *Sustainability of Water Resources under Increasing Uncertainty*, Rosjberg D, Boutayeb N, Gustard A, Kundzewiez Z, Rasmussen P (eds). IAHS Publication 240, International Association of Hydrological Sciences Press: Wallingford; 237–242.

Pomeroy JW, Gray DM. 1995. *Snow Accumulation*, *Relocation and Management*. National Hydrology Research Institute Science Report No. 7. Minister of Environment: Saskatoon. 144 pp.

Pomeroy JW, Jones HG. 1996. Wind-blown snow: sublimation, transport and changes to polar snow. In *Chemical Exchange between the Atmosphere and Polar Snow*, Wolff E, Bales RC (eds). NATO ASI Series I 43, Springer-Verlag: Berlin; 453–489.

Pomeroy JW, Li L. 1997. Development of the Prairie Blowing Snow Model for application in climatological and hydrological models. *Proceedings of the Western Snow Conference* 65: 186–197.

Pomeroy JW, Marsh P, Jones HG, Davies TD. 1995. Spatial distribution of snow chemical load at the tundra–taiga transition. In *Biogeochemistry of Seasonally snow-covered Catchments*, Tonnessen KA, Williams MW, Tranter M (eds). IAHS Publication 228. International Association of Hydrological Sciences Press: Wallingford; 191–206.

Pomeroy JW, Granger RJ, Pietroniro A, Elliott JE, Toth B, Hedstrom N. 1997a. *Hydrological Pathways in the Prince Albert Model forest: Final Report*. NHRI Contribution Series No. CS-97007, National Hydrological Research Institute: Saskatoon; 153 pp. plus appendices.

Pomeroy JW, Marsh P, Gray DM. 1997b. Application of a distributed blowing snow model to the Arctic. *Hydrological Processes* **11**: 1451–1464.

Pomeroy JW, Gray DM, Shook KR, Toth B, Essery RLH, Pietroniro A, Hedstrom N. 1998a. An evaluation of snow accumulation and ablation processes for land surface modelling. *Hydrological Processes*, **12**: 2339–2367.

Pomeroy JW, Parviainen J, Hedstrom N, Gray DM. 1998b. Coupled modelling of forest snow interception and sublimation. *Hydrological Processes* **12**: 2317–2337.

Pomeroy JW, Brown T, Kite G, Gray DM, Granger RJ, Pietroniro A. 1998c. *PBS-SLURP Model*. NHRI Contribution Series No CS-98003, National Water Research Institute, Environment Canada: Saskatoon; 25 pp. plus appendices.

Pomeroy JW, Essery R, Gray DM, Shook KR, Toth B, Marsh P. 1999a. Modelling snow–atmosphere interactions in cold continental environments. In *Interactions between the Cryosphere*, *Climate and Greenhouse Gases*, Tranter M (ed.). IAHS Publication 256. International Association of Hydrological Sciences Press: Wallingford; 91–102.

Pomeroy JW, Hedstrom N, Parviainen J. 1999b. The snow mass balance of Wolf Creek. In *Wolf Creek Research Basin*, *Hydrology*, *Ecology*, *Environment*, Pomeroy J, Granger R (eds). National Water Research Institute, Minister of Environment: Saskatoon; 15–31.

Pomeroy JW, Davies TD, Jones HG, Marsh P, Peters NE, Tranter M. 1999c. Transformations of snow chemistry in the boreal forest: accumulation and volatilization. *Hydrological Processes* **13**: 2257–2273.

Prowse TD, Ommanney CSL. 1990. *Northern Hydrology: Canadian Perspectives*. NHRI Science Report No. 1, National Hydrology Research Institute, Environment Canada: Saskatoon; 308 pp.

Quinton WL, Marsh P. 1998a. The influence of mineral earth hummocks on subsurface drainage in the continuous permafrost zone. *Permafrost and Periglacial Processes* **9**: 213–228.

Quinton WL, Marsh P. 1998b. Meltwater fluxes, hillslope runoff and stream flow in an arctic permafrost basin. *Proceedings*, *Seventh International Conference on Permafrost*, Yellowknife, NWT, Centre d'ëtudes nordiques, Laval University; 921–926.

Quinton WL, Marsh P. 1999. A conceptual framework for runoff generation in a permafrost environment. *Hydrological Processes* **13**: 2563–2581.

Reedyk S, Woo MK, Prowse TD. 1995. Contribution of icing ablation to streamflow in a discontinuous permafrost area. *Canadian Journal of Earth Science* **32**: 13–20.

Rouse WR. 2000. Progress in hydrological research in the Mackenzie GEWEX study. *Hydrological Processes* **14**: 1667–1685.
Shook K. 1995. *Simulation of the ablation of Prairie snowcovers*. PhD thesis, University of Saskatchewan: Saskatoon; 191 pp.
Shook KR, Gray DM. 1996. Small-scale spatial structure of shallow snowcovers. *Hydrological Processes* **10**: 1283–1292.
Shook KR, Gray DM. 1997a. The role of advection in melting shallow snowcovers. *Hydrological Processes* **11**: 1725–1736.
Shook KR, Gray DM. 1997b. Ablation of shallow snowcovers. In *International Symposium on Physics, Chemistry and Ecology of Seasonally-frozen Soils*. Iskandar I, Wright E, Sharratt B, Groenvelt P, Hinzman L (eds). CRREL Special Report 97–10, US Army cold Regions Research and Engineering Laboratory: Hanover, NH; 280–286.
Steppuhn H, Waddington J. 1996. Conserving water and increasing alfalfa production using a tall wheatgrass windbreak system. *Journal of Soil Water Conservation* **51**: 439–445.
Stewart RE, Leighton HG, Marsh P, Moore GWK, Ritchie H, Rouse WR, Soulis ED, Strong GS, Crawford RW, Kochtubajda B. 1998. The Mackenzie GEWEX study: the water and energy cycles of a major North American river basin. *Bulletin of the American Meterological Society* **79**: 2665–2683.
Taylor CM, Harding RJ, Pielke Sr RA, Vidale PL, Walko RL, Pomeroy JW. 1998. Snow breezes in the boreal forest. *Journal of Geophysical Research* **103**(D18): 23 087–23 101.
Tonnessen KA, Williams MW, Tranter M (eds). 1995. *Biochemistry of Seasonally Snow-Covered Catchments, Proceedings, Boulder, Colorado, July 1995*. IAHS Publication 228. International Association of Hydrological Sciences: Wallingford; 465 pp.
Van Bochove E, Jones HG, Pelletier F, Prévost D. 1995. N_2O, CO_2 and CH_4 concentrations in soil and over-lying snow cover. *Proceedings, Symposium on Trace Greenhouse Gases by Soil, Canadian Society of Soil Science*, Québec, 24–25 July.
Van Bochove E, Jones HG, Pelletier F, Prévost D. 1996. Emission of N_2O from agricultural soil under snow cover: a significant part of the N budget. *Hydrological Processes* **10**: 1545–1549.
Verseghy DL. 1991. CLASS — a Canadian land surface scheme for GCMs, I, soil model. *International Journal of Climatology* **11**: 111–133.
Waddington JM, Griffis TJ, Rouse WR. 1998. Northern Canadian wetlands: net ecosystem CO_2 exchange and climatic change. *Climatic Change* **40**: 267–275.
Webb EK. 1970. Profile relationships: the log–linear range and extension to strong stability. *Quarterly Journal of the Royal Meteorology Society* **96**: 67–90.
Woo MK. 1998. Arctic snow cover information for hydrological investigations at various scales. *Nordic Hydrology* **29**: 245–266.
Woo MK, Xia ZJ. 1996. Effects of hydrology on the thermal conditions of the active layer. *Nordic Hydrology* **27**: 129–142.
Woo MK, Young KL. 1997. Hydrology of a small drainage basin with polar oasis environment. *Permafrost and Periglacial Proceedings* **8**: 257–277.
Woo MK, Young KL. 1998. Characteristics of patchy wetlands in a polar desert environment, Arctic Canada. *Proceedings, Seventh International Conference on Permafrost*, Yellowknife, NWT, Centre d'êtudes nordiques, Laval University; 1141–1146.
Woo MK, Yang DQ, Young KL. 1999. Representativeness of arctic weather station data for the computation of snowmelt in a small area. *Hydrological Processes* **13**: 1859–1870.
Yang DQ, Woo MK. 1999. Representativenes of local scale snow data for large-scale hydrological investigations. *Hydrological Process* **13**: 1977–1988.
Young KL, Woo MK, Edlund SA. 1997. Influence of local topography soil and vegetation on microclimate and hydrology at a High Arctic site. *Arctic and Alpine Research* **29**: 270–284.
Zhao L-T, Gray DM. 1997a. A parametric expression for estimating infiltration into frozen soils. *Hydrological Processes* **11**: 1761–1775.
Zhao L-T, Gray DM. 1997b. Estimating snowmelt infiltration into medium and fine-textured, frozen soils. In *International Symposium on Physics, Chemistry and Ecology of Seasonally-frozen Soils*, Iskandar I, Wright E, Sharratt B, Groenvelt P, Hinzman L, (eds). CRREL Special Report 97–10, US Army Cold Regions Research and Engineering Laboratory: Hanover, NH; 280–286.
Zhao L-T, Gray DM, 1999. Estimating snowmelt infiltration into frozen soils. *Hydrological Processes* **13**: 1827–1842.
Zhao L-T, Gray DM, Male DH. 1997. Numerical analysis of simultaneous heat and mass transfer during infiltration into frozen ground. *Journal of Hydrology* **200**: 345–363.

