HYDROLOGICAL PROCESSES
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Advances in Canadian forest hydrology, 1995–1998

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Abstract:

Approximately 42% of Canada is covered by forests, which in turn can be subdivided into nine distinct forest ecozones. Many forested ecozones are located in northern Canada, where cold winters and cool summers provide forest environments that are less well-understood that those in more temperate locations. A number of major developments in recent years have stressed the need for enhanced understanding of hydrological processes in these forest landscapes. These include an increased emphasis on sustainable forest management in Canada as well as major scientific initiatives (e.g. BOREAS) examining water, carbon and energy fluxes in forest ecosystems, with a particular focus on boreal and subarctic forests. Recent progress in our understanding of forest hydrology across Canada is reviewed. Studies of hydrological processes across the spectrum of forest ecozones are highlighted, as well as work on hydrological responses to forest disturbance and recovery. Links between studies of hydrological processes in Canada's forests and other fields of research are examined, with particular attention paid to ongoing efforts to model hydrological impacts and interactions with the climate, biogeochemistry, geomorphology and ecology of forested landscapes. Copyright © 2000 John Wiley & Sons, Ltd.

KEY WORDS forest hydrology; Canada; interception; snowmelt; evaporation; infiltration; runoff; streamflow generation; hydrochemistry; forest disturbance; hydroecological modelling

INTRODUCTION

Canada has *c*. 10% of the world's forest cover and approximately 42% of Canada is mantled by forest. Hydrological processes in forested landscapes are of great interest, given the areal extent of forests in Canada, the recent emphasis placed on sustainable forest management in Canada, and the relevance of hydrological and hydrochemical processes to several criteria and indicators that are currently used to evaluate sustainable forestry. Of the major terrestrial ecozones in Canada, nine are forest ecozones: Taiga Shield, Boreal Shield, Atlantic Maritime, Mixedwood Plains, Boreal Plains, Taiga Cordillera, Boreal Cordillera, Pacific Maritime and Montane Cordillera (Figure 1).

The location and extent of these forest ecozones suggest that there may be dramatic differences in hydrological and hydrochemical processes between them. For example, fog drip may be a major water input in the Pacific Maritime and Atlantic maritime ecozones but of limited concern elsewhere. Conversely, an understanding of forest hydrology in the Boreal Cordillera ecozone requires a knowledge of the controls on the spatial distribution of permafrost conditions that is unnecessary when examining forests in the Mixedwood Plains. In addition to these interzone differences in the presence and relative importance of various processes and conditions, the cold winters and cool summers that affect many or all of these ecozones influence hydrological conditions in ways that are less well-understood than the hydrology of more temperate

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Figure 1. Terrestrial ecozones of Canada (after Canadian Council of Forest Ministers 1997)

locations. Thus, there is a need to consider the nature of hydrological processes and conditions that are specific to the Canadian environment.

This paper reviews recent progress in our understanding of forest hydrology across Canada for the period 1995–1998. Such an examination is timely, because the last major review of the role of forests and forest management operations in relation to water in Canada (Hetherington, 1987) occurred more than a decade ago. We present studies of a range of hydrological processes across the spectrum of forest ecozones, such that recent progress in the study of interception, evaporation, infiltration and soil water dynamics, runoff production, streamflow generation and hydrochemical fluxes in forest ecosystems will be examined. We include a review of work on hydrological responses to forest disturbance and recovery in Canada and ongoing efforts to model forest hydrology. Throughout, we emphasize the links between studies of hydrological processes and other fields of research in Canada's forests.

INTERCEPTION

Snowfall

Snow accumulation in forests prior to spring melt is a balance between snowfall, snow canopy storage as intercepted load, and ablation of canopy storage via sublimation, melt and unloading to the surface. Snow interception (that snow which has not reached the ground) varies between canopy types (Figure 2). By the end of the season, canopies are snow-free and any snow still 'intercepted' can be presumed to have sublimated. Forest canopies also influence the variability of snow water equivalent (SWE) on the ground. Pomeroy *et al.* (1998a) found coefficients of variation of SWE to increase with evergreen canopy density in the boreal forest, but Faria (1998) found greatest SWE variability under canopies of medium density and in mixed-wood forests.

A modification (Pomeroy and Gray, 1995) of an empirical forest snow depth equation developed in British Columbia coastal forests by McNay *et al.* (1988) enabled estimation of SWE as a linear function of evergreen canopy coverage and above-canopy snowfall. This empirical formulation worked reasonably well for boreal forests if snowfall amounts were small (Hedstrom, 1998). Subsequent research in the boreal forest yielded a

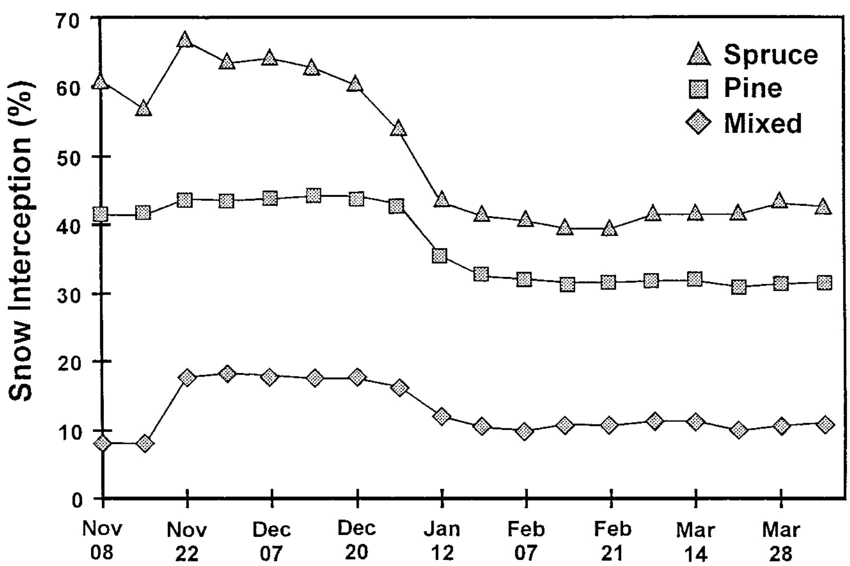

Figure 2. Snow interception (measured as the difference between above- and below-canopy areal snowfall) shown as a percentage of cumulative seasonal snowfall from sites near Waskesiu, Saskatchewan, 1994–95 (after Pomeroy *et al.*, 1998b)

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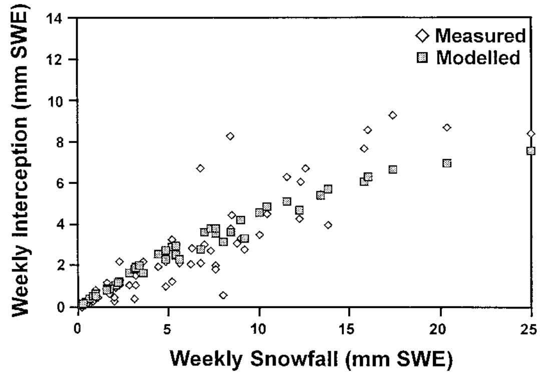

Figure 3. Measured (as determined from above- and below-canopy snowfall) and modelled snow interception (1994–97) for a pine canopy in central Saskatchewan (after Hedstrom and Pomeroy, 1998)

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). Modelled interception showed good correspondence with measured weekly interception (Figure 3). The model suggested 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 much larger than rainfall interception and much greater than the parameterizations of snow interception presently used by land surface schemes such as CLASS (Pomeroy *et al.*, 1998a). Hedstrom and Pomeroy (1998) proposed that the rate of unloading of intercepted snow was an exponential function of time but were unable to test this hypothesis directly with field measurements.

Sublimation reduces the amount of intercepted snow 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. As intercepted snow does not increase the albedo of evergreen canopies significantly (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 interception (Hedstrom and Pomeroy, 1998) and snow sublimation algorithms (Pomeroy and Gray, 1995) to calculate canopy snow mass balance and surface snow accumulation. Modelled values compared well with measurements (Figure 4) and suggested that sublimation can ablate up to 5 mm SWE per day in the southern boreal forest, resulting in relatively rapid loss of forest snow.

Rainfall

Price *et al.* (1997) examined water fluxes through a spruce canopy and moss understorey in northern Manitoba during rainfall. Stemflow was insignificant but the canopy intercepted 15–60% of rainfall, interception decreasing with storm quantity. Over the summer season, canopy interception comprised 23% of rainfall, an amount almost doubled when moss interception was included. Moss interception accounted for 23% of throughfall to the ground, with subsequent evaporation comprising 21% of total moss interception. On days with large moss interception, evaporation of intercepted rainfall from moss comprised about 50% of the total evaporative flux.

In a hardwood stand in southern Ontario, Carlyle-Moses and Price (1999) evaluated two versions of the Gash rainfall interception model (Gash, 1979; Gash *et al.*, 1995). Stemflow and canopy interception comprised 4% and 19% of rainfall, respectively. Both original and revised Gash models simulated the observations within the range of measurement error; however, the authors recommended the latter model because

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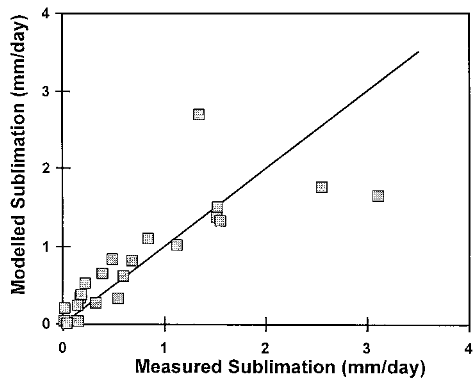

Figure 4. Canopy snow sublimation modelled using the Pomeroy *et al.* (1998b) model and measured using the weight of snow on a suspended, weighed tree. The weight of snow on the tree was converted to an areal SWE using comparative snow surveys during accumulation periods

of its improved physical basis. Of the interception loss, the model indicated that roughly 30% evaporated during rainfall and 60% evaporated afterwards.

Pomeroy *et al.* (1997) implemented a form of the Rutter (Rutter *et al.*, 1975) interception model, using a modification of Granger's (Granger and Gray, 1989; Granger and Pomeroy, 1997) evapotranspiration model to remove water from the canopy. The model was applied to open and closed canopy locations in a mixed-wood of aspen and spruce, a mature pine and a regenerating pine canopy. Evaporation of intercepted rainfall as a percentage of total summer evapotranspiration ranged from 3% in an open section of the regenerating pine stand to 37% under spruce trees in the mixed-wood. Partitioning of summer rainfall between drip, canopy evaporation and throughfall varied with stand type and canopy density (Figure 5).

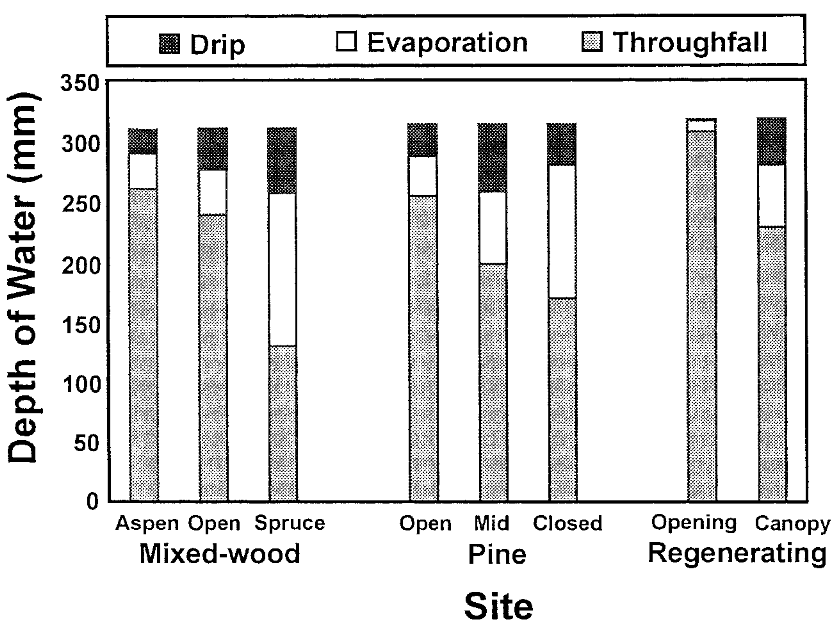

Figure 5. Partitioning of summer rainfall in a Saskatchewan boreal forest — values determined from measurements and modelling (after Pomeroy *et al.*, 1997)

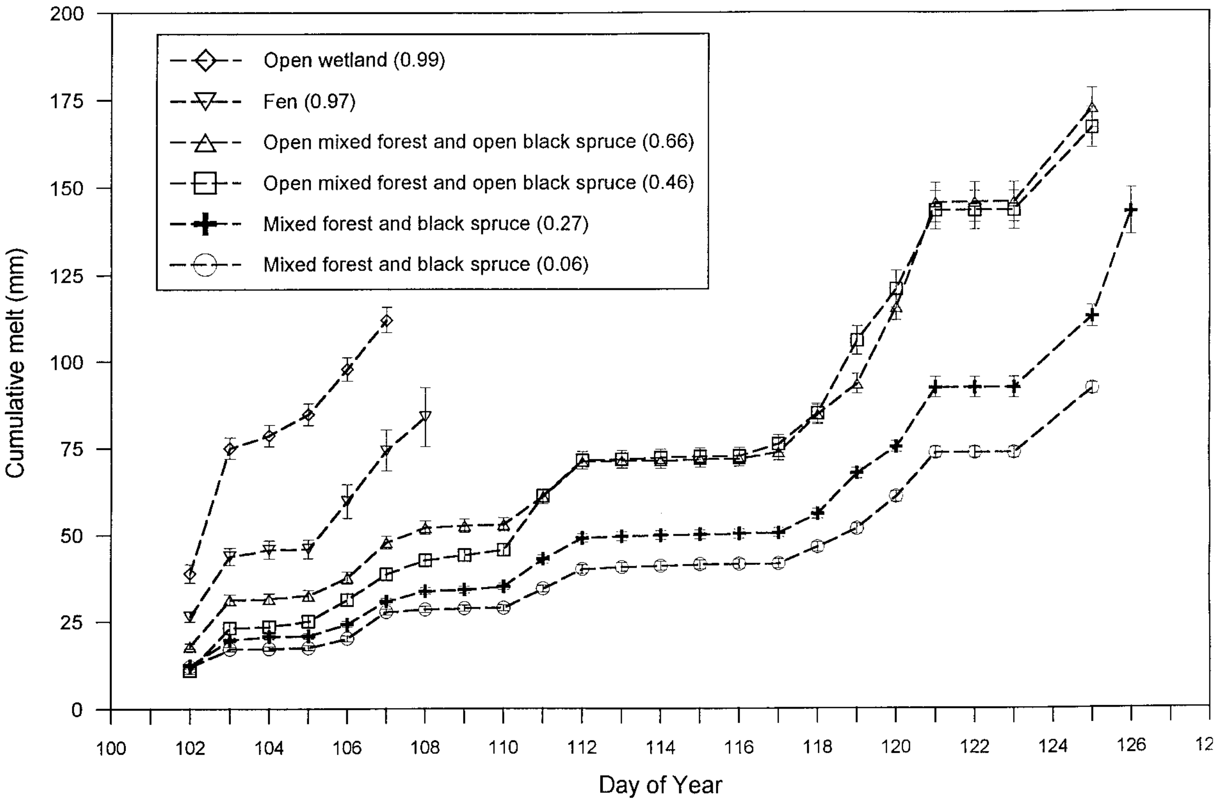

Figure 6. Cumulative melt measured at various sites in the boreal forest of northern Manitoba, 1994. Error bars indicate ± 1 standard error about the mean melt depth. Values in parentheses are the canopy gap fractions (fraction of visible sky) at each site (after Metcalfe and Buttle 1998)

Snowmelt

Snowmelt under forest canopies is strongly altered from the open environment case because the overlying canopy intercepts radiation and suppresses turbulent transfer. Melt rates are thus slower under forests than in equivalent open areas, a factor used in water management for many decades. Pomeroy and Dion (1996) examined the role of canopy leaf area in extinguishing short-wave radiation, and the relationship between this extinguished short-wave and subcanopy net radiation during snowmelt. Daily mean solar angle exerted a strong control on light extinction in pine canopies, and therefore on subcanopy net radiation. Subcanopy net radiation in a medium density pine canopy in central Saskatchewan was not positive until late March, and its magnitude was controlled by solar-angle and insolation rather than air temperature. Metcalfe and Buttle (1997, 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. An exponential relationship between snowmelt rate and non-canopy coverage (gap fraction) was used to up-scale and map snowmelt over a heterogeneous area, with canopy coverage obtained from classified Landsat images. An example of variable depletion of snow cover by snowmelt under different canopy types in northern Manitoba is shown in Figure 6.

Pomeroy and Granger (1997) measured melt energetics in stands of mature pine, mixed aspen–spruce, regenerating pine and cut-block in a Saskatchewan boreal forest. Subcanopy radiation dominated the snowmelt energy balance, with convective terms tending to cancel out and become small in all environments. The largest net convective terms were measured under mature pine canopy and accounted for 22% of total

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snowmelt energy during melt. Convective terms contributed to melt for the first 10 days in the pine canopy, the first 5·5 days in the mixed-wood, the first 2·5 days in the regenerating pine and for only the first few hours in the cut-block out of a total snowmelt period of 15 days. A threefold increase in snowmelt rate upon removal of a mature boreal forest was inferred from this study and the complete forest hydrology report (Pomeroy *et al*., 1997).

Faria (1998) investigated the influence of forest canopy on snow cover depletion. Depletion was greatly delayed by energy flux modifications in dense canopies, although the rate of depletion was advanced somewhat because of the covariance of melt energy and SWE. This covariance was strongest under mixed-wood and medium density canopies. The effect of topography on forest snowmelt was investigated by Carey and Woo (1998), who compared snowmelt on an open spruce slope (north facing) to a dense aspen forest (south facing) in a Yukon valley. Snowmelt was advanced by 10 days on the south-facing slope.

EVAPORATION

Although evaporation and transpiration (henceforth combined and referred to simply as evaporation, E) are major components of the hydrological cycle in Canadian forests, evaporative processes are poorly understood, particularly in northern areas of Canada (Prowse, 1990). Several recent studies related to the evaporation of intercepted water have been described above. In addition, a number of important studies conducted during the 1995–98 period have led to significant improvements in our understanding of evaporation in Canada's forests. One of these is the Mackenzie basin GEWEX Study (MAGS), which is the Canadian contribution to the Global Energy and Water Experiment (GEWEX). Those aspects of MAGS that relate to evaporation from forest cover in the Mackenzie basin are addressed by Rouse (2000, this issue). A concurrent major study that deserves mention, and that also focused extensively on evaporation from forest landscapes, was the Boreal Ecosystem–Atmosphere Study (BOREAS) conducted in the Boreal Shield ecozone of central Canada.

The BOREAS project

BOREAS was an attempt to improve our understanding of interactions between the boreal forest biome and the global climate system (McCaughey *et al*., 1997). A major component of this project was the study of surface–atmosphere exchanges of water, with an emphasis on evaporation. Study sites in central Saskatchewan and northern Manitoba were used. BOREAS represents the most intensive set of studies of forest evaporation in Canada during the 1995–98 period, combining flux tower estimates, flux aircraft measurements and satellite remote sensing. Initial results (Sellers *et al*., 1995) revealed surprisingly small E during the growing season, arising from two factors.

1. Soil properties. The root zone for conifers in this portion of the Boreal Shield ecozone is less than 0·4 m, and is often underlain by low-permeability substrate (McCaughey *et al*., 1997). Thus, water inputs run off rather than enter soil water storage for subsequent abstraction by E.
2. Physiological control exerted by trees in response to large vapour pressure deficits. This leads to small transpiration rates from the various tree species that comprise the boreal forest.

A more complete presentation of the major findings of BOREAS was provided in a special issue of the *Journal of Geophysical Research*, and the reader is referred to this for the results of individual research projects. An overview of the individual papers was given by Sellers *et al*. (1997), who noted that the small evaporation rates observed at BOREAS forest sites (particularly the coniferous covers) are not represented correctly in most atmospheric models. Root system are frozen during the spring and canopy transpiration is cut off. Delayed soil thawing relative to more temperate forests (owing to interception and partitioning of most of the radiation flux by the forest canopy) results in very large sensible heat fluxes over the forest. Evaporation was depressed during the middle of the growing season owing to the small stomatal conductances associated minor photosynthetic rates of these species, which are adapted to nutrient-poor

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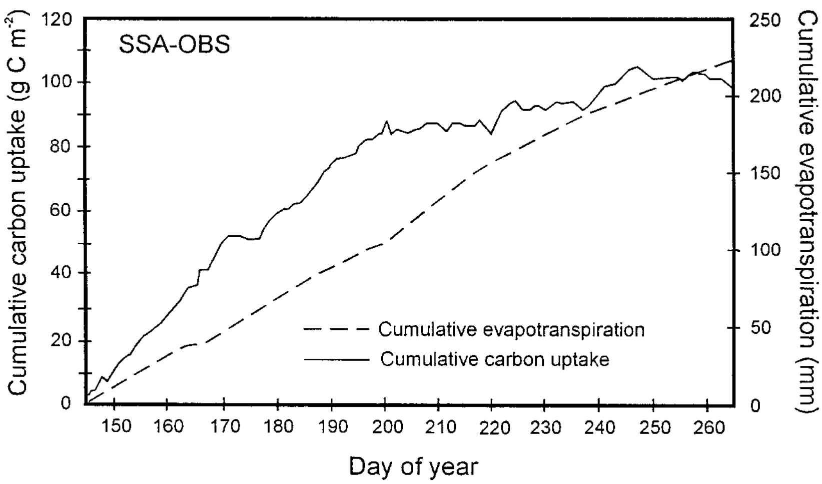

Figure 7. Cumulative carbon uptake and evapotranspiration from measurements made above a black spruce canopy (from Jarvis *et al.*, 1997)

environments. Flux aircraft measurement of mesoscale E agreed with tower flux measurements showing much smaller E than was previously thought, although the flux aircraft seemed to underestimate systematically the surface fluxes. Aircraft-measured E fluxes were found to be closely related to spectral vegetation indices, and several studies observed a close link between E, photosynthesis and carbon uptake (Figure 7).

Other studies of forest evaporation

Lafleur and Rouse (1995) reported on five years of summertime energy budgets for continuous wetland tundra and treeline forest near Churchill, Manitoba. Evaporation was consistently less at the forest site relative to the tundra site. Wind direction relative to Hudson Bay was used as an analogue for changing climatic conditions, where onshore winds are cooler and more moist than offshore winds. Lafleur and Rouse (1995) contended that as the climate warms and becomes drier, more additional energy goes into E from the wetland tundra than from the forest.

Pomeroy *et al.* (1997) studied E from four sites (mature jack pine stand, mixed aspen and white spruce stand, regenerating site and cut-block site) in the boreal forest in Prince Albert Model Forest (PAMF), Saskatchewan. An extension of the Penman model for E from a saturated surface (developed by Granger and Gray, 1989) was used to model E based on micrometeorological measurements (net radiation, air temperature, wind speed and humidity). Estimates were compared with daily E measured using eddy-correlation equipment deployed for selected periods in cut-block and regenerating sites. There was excellent agreement between estimated and measured cumulative E for both sites, and the model was then used to determine E for all four sites for the May–September period for three years. The greatest E in each year was from the pine stand. Evaporation from other stands ranged from 0·87–0·88 (mixed-wood), 0·74–0·81 (regenerating) and 0·69–0·71 (cut-block) of E from the pine stand.

Pomeroy *et al.* (1997) also determined Bowen ratios (ratio of sensible heat flux to latent heat flux) for the sites. The Bowen ratio at the cut-block site remained relatively constant (near unity) throughout the study, indicating that the site could not exert a control on energy partitioning and thus reduce E during drier periods. Mature forest stands had smaller Bowen ratios, indicating that a greater portion of received energy was used in evaporation rather than surface heating. Bowen ratios at mature stands also increased during drier periods, reflecting the forest's ability to control evaporative losses. Over a three-year period, Bowen ratios from the regenerating stand showed a transition from the relatively constant values exhibited by the cut-block to an increased response to changing environmental conditions, typical of mature forest stands.

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Lafleur *et al.* (1998) examined the summer micrometeorology of two forest stands located a short distance (12·8 km) apart in eastern Ontario over a three year period. The first was a primarily deciduous forest with fine sandy soils, whereas the second was a mixed forest over a thin sandy-loam soil. Latent heat fluxes (and thus E) were larger at the deciduous forest site. Differences were attributed to soil moisture stress at the mixed forest site, which restricted evaporation. Instead of observing a gradual decline in evaporation with soil moisture stress (Black, 1979), Lafleur *et al.* (1998) found an 'on–off' switching system, where a decline in soil water content to $<4{\cdot}5$ cm in the upper soil profile resulted in an abrupt decline in E from the mixed forest stand. They recommended that future work focus on the nature of this phenomenon.

INFILTRATION, SOIL WATER STORAGE AND MOVEMENT

Water entry into and storage within forest soils exert a strong control on water pathways and runoff generation in forest landscapes. Soil water storage also influences transpiration, as reductions in available water lead to development of water stress in trees (McCaughey *et al.*, 1997). Infiltration and water storage are also relevant to nutrient cycling in forests and the fate of non-point pollutants reaching the forest canopy. Recent work has focused on three general themes: infiltration into frozen soil, effects of preferential flowpaths on water movement in forest soils, and evaluation of the soil-water balance.

Infiltration into frozen soils

Infiltration is critical to the partitioning of meltwater and rainfall between overland flow and entry into soils, and infiltration studies in frozen soils must consider the key roles assumed by phase changes and thermal processes (Zhao *et al.*, 1997). Pomeroy *et al.* (1997) used a twin-probe gamma system to measure total soil moisture and soil moisture changes during snowmelt at sites (mature jack pine stand, mixed aspen and white spruce stand, regenerating site and cut-block site) in the boreal forest at PAMF. Soil temperatures were measured and infiltrometer tests were used to determine 'final' infiltration rates for the sites. Snow covers at the cut-block site were often $>2\times$ the snow depth under coniferous canopies. This deeper snow cover provided more insulation to cleared soils, which were further warmed by early partial melts. As a result, the upper soil profile at the cut-block site was thawed during the snowmelt period, whereas the upper soil under coniferous canopies was still frozen. During periods of significant snowmelt, the soil temperatures at the various sites were not very low ($-3{\cdot}8$ to $-0{\cdot}3\,^{\circ}$C) and initial soil water contents of near-surface soil were small.

Pomeroy *et al.* (1997) noted that all SWE generally infiltrated if water entry into forest soils was not impeded at the surface by an ice lens or other obstruction, and if the average initial soil-water/ice content of the 0–1 m soil layer was $<0{\cdot}14$ m^3 m^{-3}. These observations were used to derive an empirical equation for infiltration into the upper 1 m of forest soils (I, mm) during snowmelt

$$I = (1 - 1{\cdot}36\theta_{\mathrm{p}})\frac{\mathrm{SWE}^{2{\cdot}41}}{364} \tag{1}$$

where θ_{p} is the degree of pore saturation (water content divided by porosity) of the upper 1 m of soil prior to snowmelt and SWE is snow water equivalent of snow cover plus any precipitation during snowmelt (mm). The model provided a reasonable fit to measured infiltration, and shows considerable promise as a means of producing realistic estimates of infiltration to frozen soils.

Zhao *et al.* (1997) and Zhao and Gray (1997) used a numerical model (HAWTS, heat and water transport in frozen soils) to study the physics and mechanics of infiltration in frozen soils. The model was based on field evidence showing a rapid decrease in infiltration rate with time immediately after the start of infiltration into frozen soils. This was followed in turn by a quasi-steady state regime in which the infiltration rate decreased slowly with time, similar to unfrozen soils. They found the following general relationships between cumulative infiltration into frozen soils (INF) and soil properties

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$$\mathrm{INF} \propto S_0 \tag{2}$$

$$\mathrm{INF} \propto (1 - S_1) \tag{3}$$

$$\mathrm{INF} \propto \frac{1}{\left(\frac{273{\cdot}15 - T_1}{273{\cdot}15}\right)} \tag{4}$$

$$\mathrm{INF} \propto t^{0{\cdot}5} \tag{5}$$

whereas S_0 is initial surface saturation, S_1 is initial soil saturation (liquid water plus ice), T_1 is initial soil temperature (K) and t is time. Model results are consistent with previous theoretical and empirical studies of infiltration. Thus, Equation (2) reflects increasing infiltration with increasing capillary pressure gradient and relative permeability at the surface, and Equation (3) indicates a direct relationship between infiltration and air-filled porosity prior to infiltration. Equation (4) agrees with the observed decrease in infiltration with decreasing initial soil temperature (colder soils have smaller unfrozen liquid water contents and relative conductivities). Finally, Equation (5) has also been observed for unfrozen soils where infiltration is dominated by the capillary pressure gradient.

Preferential flow in forest soils

Forest soils contain abundant macropores (pore spaces significantly larger than those of the soil matrix), particularly in near-surface soil as a result of large root densities and faunal activity. Although they make up a relatively small fraction of the soil's total porosity, macropores can have a disproportionate effect on the soil's infiltration properties, allowing infiltrating water to bypass the soil matrix and reach specific depths ahead of water moving via soil micropores. Buttle and House (1997) combined field (single ring infiltrometer) and laboratory (constant head permeameter) measurements of saturated hydraulic conductivity (K_H) to examine effects of macropores on K_H in a small forest basin with shallow soils in south-central Ontario. The value of K_H for macroporous soil was similar to that of the soil matrix for porous hillslope podzols; however, it could be several orders-of-magnitude greater than the matrix K_H for gleysols in the stream valley. Geostatistical analyses revealed that the degree of anisotropy in the spatial distribution of K_H and the scale of spatial dependence were smaller for the bulk profile K_H (which includes the effects of macropores) relative to the minimum matrix K_H values in the profile at a site.

Soil-water balances

Water storage in forest soils has important implications for a number of hydrological processes, including evaporation and runoff generation. In addition, available soil moisture may limit reforestation of cut-block sites (Elliott *et al*., 1998). There have been two major recent studies of soil-water storage in the boreal forest. The first was that of Cuenca *et al*. (1997), who measured root zone soil-water content and soil hydraulic properties at BOREAS flux tower sites in central Saskatchewan and northern Manitoba. Vertical profiles of total hydraulic potential were obtained by applying estimated soil-water retention functions to soil-water content measurements, and the position of the zero-flux plane (separating E from drainage) was determined. Cuenca *et al*. (1997) used the soil water balance (SWB) approach to estimate E and drainage, and compared their results with E from tower fluxes. They found good agreement between SWB and tower flux estimates of E during the late spring; however, the SWB produced much smaller E values during mid- and late-summer periods relative to tower flux estimates.

Elliott *et al*. (1998) measured soil water profiles at two mature forest sites (pine and mixed-wood) and two plantations (pine and mixed wood) at PAMF, along with precipitation, interception and evaporation. Variability in moisture storage within stands was associated with canopy structure and density, water extraction patterns and mechanical site preparation. Changes in soil-water storage within the upper 1 m agreed well with micrometeorological estimates of E. Elliott *et al*. (1998) also used a water balance approach

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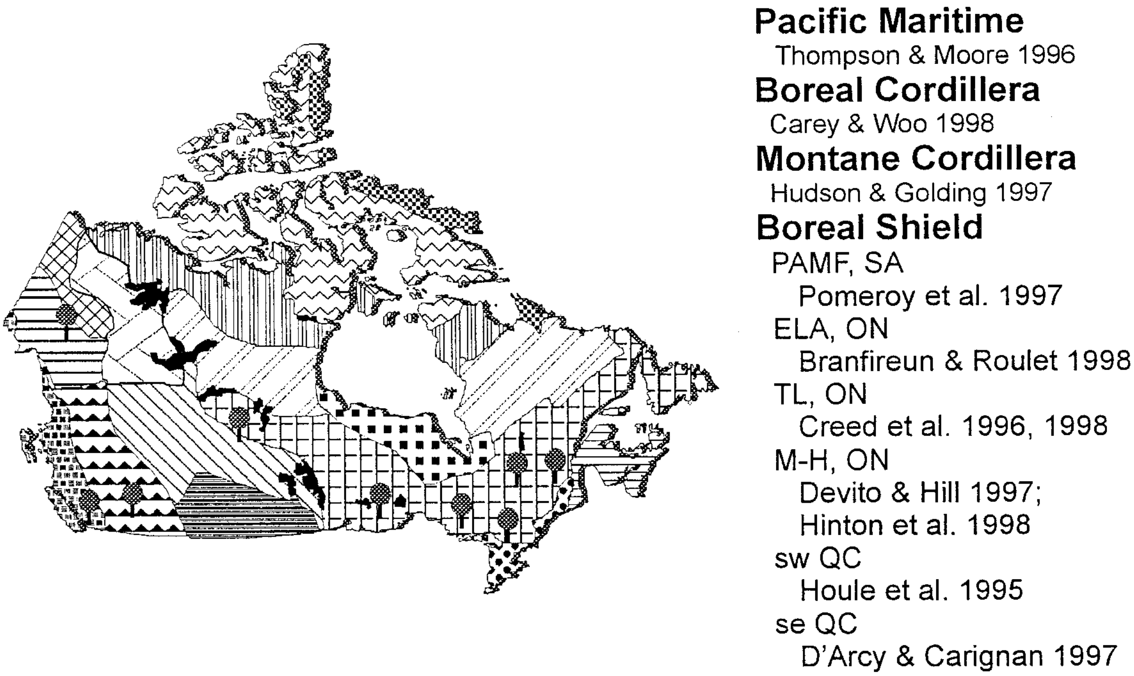

Figure 8. Locations and examples of recent studies of runoff and streamflow generation and/or their implications for hydrochemical fluxes in various forest ecozones

to estimate runoff response to a >100 mm mid-summer storm for all sites. There was no runoff from mature forest stands but 19–35 mm of runoff was estimated for regenerating sites, with greater runoff from the more recent plantation.

RUNOFF PROCESSES, STREAMFLOW GENERATION AND HYDROCHEMICAL FLUXES

Figure 8 indicates sites where studies of runoff and streamflow processes and/or their implications for hydrochemical fluxes in forested landscapes have been undertaken during the 1995–98 period. It demonstrates the uneven distribution of studies across Canada's forest ecozones, and highlights the focus of such studies in the southern portion of the Boreal Shield ecozone.

The forested Precambrian Shield

Several studies conducted in the Ontario portion of the Boreal Shield ecozone stress the role that the relative thickness and spatial uniformity of overburden play in determining: (i) the role of various runoff processes in forested basins on the Precambrian Shield; and (ii) the degree of coupling between hillslopes and receiving waters (e.g. streams, wetlands, lakes) and its influence on basin streamflow characteristics (Figure 9). This framework also assists in synthesizing work on hydrochemical fluxes in forested basins conducted at the slope and basin scales in these landscapes.

Runoff processes

Comparison of landscapes with varying degrees of overburden thickness and uniformity reveals the strong control that the overburden mantle can exert on the type and relative importance of various runoff processes at work on the Precambrian Shield. Slopes in the Experimental Lakes Area (ELA) of north-western Ontario consist of a patchwork of bedrock outcrops and soil 'islands' that support forest stands. Bedrock outcrops

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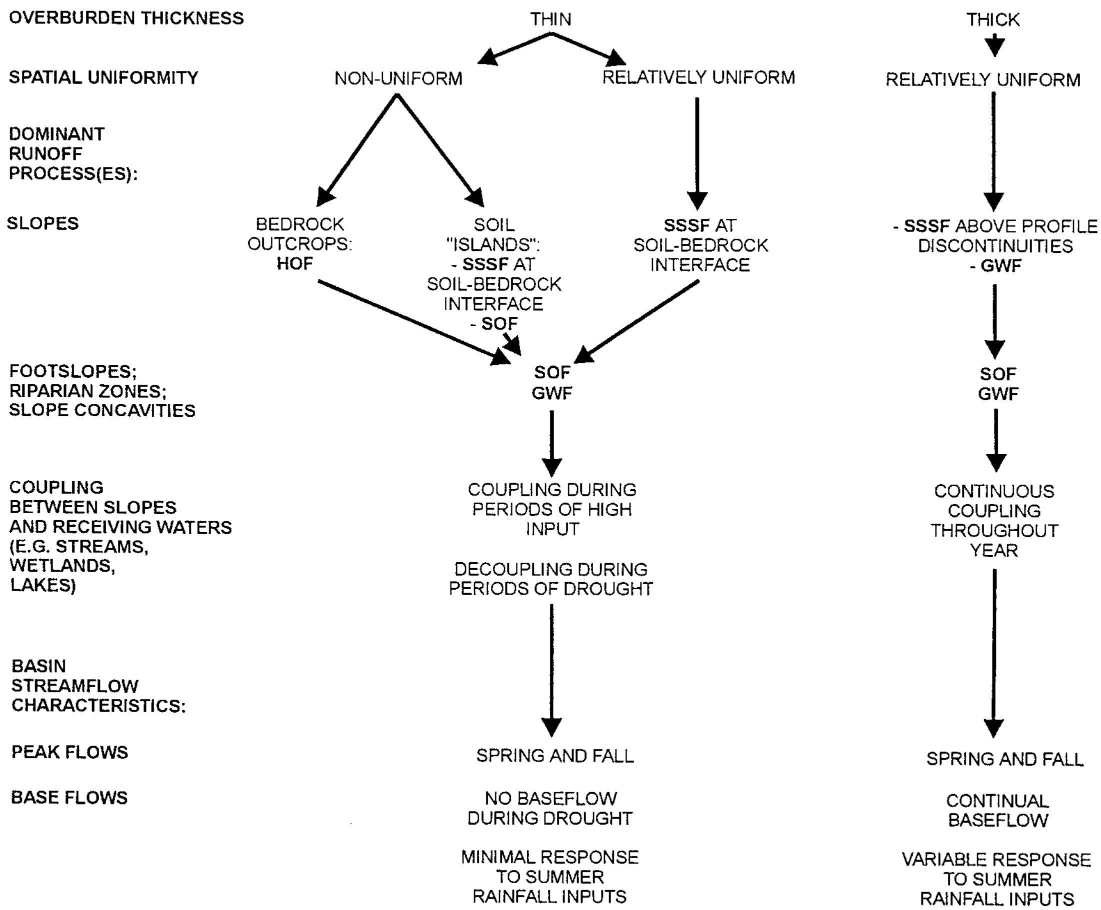

Figure 9. Generalized influence of overburden thickness and spatial uniformity on runoff processes, hydrological coupling between slopes and receiving waters, and basin streamflow characteristics

produce Horton overland flow (HOF) during rainfall and snowmelt. Runoff from the forest stands occurs mainly as subsurface stormflow (SSSF) through surface organic LFH soils, although saturation overland flow (SOF) can be generated when large inputs, combined with appropriate bedrock topography beneath the forest soil, result in complete saturation of the soil profile (Allan and Roulet, 1994). Slope runoff in areas of thin (<1 m) but relatively uniform overburden appears to occur as SSSF. Thus, studies of slope hydrology near Dorset, south-central Ontario (Peters *et al.*, 1995; Buttle and Peters, 1997), revealed that subsurface flow was concentrated in a thin zone above the soil–bedrock interface, and that this process supplied $>90\%$ of total slope runoff during rainfall events. Water inputs to the slope surface reached the bedrock as preferential flow through soil macropores, as indicated by reactive (dissolved silica) and non-reactive (oxygen-18) tracers. Conversely, slopes underlain by relatively thick overburden (up to 15 m) in the Muskoka region of south-central Ontario supply runoff to the riparian zone largely as groundwater flow (GWF) moving through the glacial till (Hinton *et al.*, 1994). Footslope areas, riparian zones and hillslope concavities tend to have greater antecedent wetness conditions and shallower water tables than further upslope (Maclean *et al.*, 1995). Thus, such sites tend to generate runoff largely by a combination of SOF and GWF (e.g. Creed *et al.*, 1996).

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Streamflow generation

The ability of forest basins to generate streamflow, or to deliver water to a receiving wetland or lake in response to a given water input, is strongly controlled by the degree to which hillslopes are coupled to the receiving water body (Figure 9). This was noted in work by Devito *et al.* (1996), who examined groundwater–surface water interactions in forested headwater wetlands. Water delivery from forested slopes to receiving wetlands was largely governed by overburden depth, such that slopes with deep overburden continued to supply groundwater to the wetland throughout the year (continuous hydrological coupling). Continuous coupling between slopes and receiving streams was also observed in a headwater basin with relatively deep overburden, such that the basin continued to deliver baseflow during the summer months (Hinton *et al.*, 1998). Devito *et al.* (1996) found that slopes with shallow overburden drained quickly following spring snowmelt, with no detectable water table in the footslope zone. This led to a decoupling between slopes and receiving wetlands during summer drought, with the result that baseflow ceased in streams draining the wetlands (Figure 9). Slopes with shallow overburden may also become decoupled from headwater streams during extended summer droughts, such that the streams dry up (Goodyear, 1997; Hinton *et al.*, 1998).

Forest basins generally produce peakflows during spring snowmelt and to a lesser extent during autumn rainstorms. Significant differences in hydrograph form, however, emerge when basins with continuous coupling between slopes and receiving waters are compared with those that experience periodic decoupling. The former produce continual baseflow and generally show short-lived increased streamflow in response to summer rainfalls. This was observed by Branfireun and Roulet (1998) in their study of the hydrological linkage between hillslopes and a peatland stream at ELA. Streams draining basins where slopes may become decoupled from receiving waters often dry up completely during prolonged summer drought, and streamflow may not respond to any but the largest rainfall inputs.

Runoff processes and streamflow generation in other forested landscapes

Studies in forest types other than those of the Boreal Shield ecozone include an examination of the snowmelt hydrology of two forested slopes in the boreal Cordillera ecozone of southern Yukon Territory (Carey and Woo, 1998). Snowmelt in aspen forest cover on a south-facing slope infiltrated the frozen silt mantle without generating runoff. Conversely, meltwater in open spruce stands on north-facing slopes occurred later, infiltrated the frozen organic soil but accumulated above the permafrost table. Water flowed laterally downslope along intermittent rills fed by diffuse and pipe flows. The work emphasized the control that aspect plays in runoff generation in subarctic forests.

Thompson and Moore (1996) studied the relationship between topography and near-surface groundwater conditions in a shallow forest soil basin in the Coast Ranges of southwestern British Columbia (Pacific Maritime ecozone). Previous research in this landscape showed that almost all water infiltrated the highly permeable soils and moved downslope as a saturated layer at the contact between soil and underlying till or bedrock (similar to the mechanism described by Peters *et al.*, 1995). The relationship between topographic characteristics derived from a raster digital elevation model (DEM) and water table depth was explored. Thompson and Moore (1996) found that the $1n(a/\tan\beta)$ index (where a is upslope contributing area and $\tan\beta$ is gradient) provided the most reliable classifications of water table elevation. They also showed how the predictive accuracy of the index varied inversely with DEM grid size, and stressed the potential application of this approach to the mapping of saturated source areas. Recognition of the dominance of water movement via subsurface pathways for runoff generation in forested mountainous basins also led Loukas and Quick (1996a) to use kinematic wave equations to model both slope runoff contributions to stream channels and channel flow in order to predict basin lag time.

Hydrochemical fluxes in forested basins

Recent work on this subject can be categorized loosely as reductionist studies of hydrological–hydrochemical linkages or basin-scale studies linking hydrological and hydrochemical processes (Dillon and Molot, 1997).

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The former are generally spatially limited plot studies or short-term process-oriented studies essential for the identification of important principles and processes, whereas the latter are useful in the evaluation of the interaction of processes over larger landscapes and longer time periods (Dillon and Molot, 1997).

Reductionist studies. Studies of hillslope hydrology and hydrochemistry have emphasized the role of water residence time in controlling solute levels in runoff reaching receiving waters. Allan (1995) examined runoff in a heterogeneous Precambrian Shield landscape at ELA, using additions of H_2SO_4 and NH_4NO_3 to snow cover prior to snowmelt. Basins dominated by bedrock outcrops experienced little biotic retention of NO_3 whereas runoff from basins containing substantial wooded soil deposits had a more protracted travel time and greater NO_3 retention. Widespread occurrence of HOF and SOF in this landscape (Figure 9) and the consequent short residence time that water has on slopes also has been cited as a possible cause of the undersaturation of runoff waters in terms of $Al(OH)_3$ and Al^{3+} relative to $Al(OH)_3$ mineral-solubility controls. Peters *et al*. (1995) and Buttle and Peters (1997) examined preferential movement of event water through forest slopes using reactive (silica) and non-reactive (oxygen-18) tracers. Dissolved organic carbon (DOC) tends to be taken up in the B horizons of these forest podzols, and preferential flow permits transport of DOC to the slope base and possibly to the receiving stream channel, lake or wetland. An extension of this work revealed some preferential flow of NO_3, NH_4 and dissolved organic nitrogen (DON) in forest slopes, although this transport was regulated by the biogeochemistry of the organic horizon of the soil (Hill *et al*., 1999).

Creed *et al*. (1996) used a topographically based hydrology model (TOPMODEL) to infer subsurface flow and saturation overland flow processes in the footslope area of forested basins. They postulated a nitrogen-flushing mechanism during spring snowmelt and autumn stormflow in a sugar maple forest in the Turkey Lakes Watershed (TLW) in central Ontario, whereby a rising water table extends into previously unsaturated parts of a N-enriched soil profile. This in turn results in surface and saturated subsurface flow that flushes N into the stream, leading to large N export (Creed and Band, 1998a). Evidence for this N-flushing mechanism was also provided by research in forested subalpine basins in southern British Columbia (Hudson and Golding, 1997).

Several studies have examined hydrochemical processes in forested landscapes in terms of the magnitude and timing of hydrological coupling of hillslopes and receiving waters. Devito and Hill (1997) found that wetlands in southcentral Ontario that become hydrologically decoupled from hillslope water inputs during summer experience water table drawdown, re-oxidation of accumulated S, and peak SO_4 concentrations in wetland stormflow outputs. Wetlands that remain hydrologically coupled to hillslopes maintain high water table conditions, experience S reduction and net retention of S. Coupling of hillslopes and a receiving peatland stream at ELA resulted in stormflow that dominated methylmercury (CH_3Hg) transport to a downstream pond (Branfireun *et al*., 1996). Hillslope–streams linkages and their implications for export of other substances to receiving waters were stressed in other studies. Thus, Maclean *et al*. (1995) observed increased H and SO_4 export from snowpack to stream during rain-on-snow inputs combined with large antecedent wetness conditions in the footslope and near-stream zones of a small basin in the Muskoka region of southcentral Ontario. Working in the same area, Hinton *et al*. (1998) found that a strong degree of coupling between slopes and stream produced higher water table levels in the riparian zone and relatively greater export of DOC to the stream than for a basin in which the stream became decoupled from the hillslope during summer drought.

Basin-scale studies. There have been several recent attempts to explain regional variations in hydrochemical fluxes to streams and lakes in terms of the general properties of the terrestrial landscape, based on the assumption that these properties are correlated with the dominant hydrological processes. Houle *et al*. (1995) examined fluxes of DOC and dissolved organic sulphur (DOS) from forested basins to lakes in southwestern Quebec. Basin properties such as altitude and flowpath influenced DOC and DOS levels in lakes as well as dissolved organic matter (DOM) supply and composition. D'Arcy and Carignan (1997) found that basin slope contributed to the explanation of water chemistry in 30 Canadian Shield lakes of south-eastern Quebec. The

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relationship between slope and chemical parameters was variable (inverse in the case of DOC, total phosphorus, chlorophyll a, Ca and Mg, direct in the case of NO_3 and NH_4). They proposed that the strong influence of basin slope on water quality is due to slope-dependent seasonal waterlogging, which determines retention or export to surface waters of dissolved substances produced within and moving through the forest floor.

Dillon and Molot (1997) examined the role of landscape composition on the export of DOC, Fe and P from forested basins in southcentral Ontario. Export was governed largely by the proportion of a basin occupied by wetlands, and the general properties of upland portions of basins were not found to explain a significant portion of variation in long-term export. St. Louis *et al.* (1996) compared CH_3Hg and total mercury (THg) dynamics in four boreal forest basins in north-western Ontario. Forested upland areas were found to be consistent sinks of CH_3Hg whereas wetlands in the basin were important sites of CH_3Hg production. Transfer of CH_3Hg from forested basins to receiving lakes may be modelled using information on the proportional wetland coverage within a basin, wetland type and annual basin water yield.

Recent research (e.g. Creed and Band, 1998a) shows that hydrochemical export patterns from contiguous basins with relatively homogeneous climate, forest and soil cover in the TLW region may be highly variable, and that this variation can be attributed largely to basin topography. This theme was addressed by Creed and co-workers in a number of papers.

Observations of significant natural variations in water, dissolved inorganic nitrogen (DIN) and DON fluxes suggested analysis of N export as a function of discharge in an effort to identify a unifying mechanism of N export (Creed and Band, 1998b). Thus, DON was found to be a linear function of discharge, indicating constant release of DON from the soil to the stream. In contrast, DIN export was an exponential function of discharge, implying that DIN release varied with hydrological conditions. A N-flushing hypothesis was formulated, whereby DIN accumulates in upper soil layers of the soil during dry periods. Formation of a saturated subsurface layer during periods of water table rise flushes DIN from the upper soil layers into the stream.

Basin controls on the N-flushing activity were also explored (Creed and Band, 1998a). A basin's flushing time was defined as the time interval that must elapse in order for N concentrations in streamflow to decline to 37% (e^{-1}) of initial values. A significant range in flushing response was observed at TLW, and basins with longer flushing times also had larger N export. Export of N was related to the incremental development of new flushing areas over the course of events. Basins with topography conducive to lateral expansion of flushing areas have longer flushing times and larger N export, whereas basins with topography that inhibits lateral expansion of flushing areas have shorter flushing times and smaller N export. The results support those of Dillon and Molot (1997), who also cited the influence of topography on hydrological processes that regulate export of dissolved nutrients from forests to surface waters.

The N-flushing hypothesis was tested using topographically based similarity indices of N formation and flushing from a distributed hydro-biogeochemical simulation model (Creed *et al.*, 1996). Covariation of these indices of N dynamics was related to observed N export among basins. The similarity indices captured N export dynamics, both within a single basin during spring melt or autumn storms and summer droughts (Creed *et al.*, 1996) and among basins representing the range of topographic variability within the forest (Creed and Band, 1998b). The approach stressed the importance of hydrological coupling between hillslopes and stream channels to N export, as demonstrated by the empirical studies of streamflow generation and hydrochemical fluxes discussed above (Figure 9). These N similarity indices provide support for the N-flushing mechanism, as well as a basin for generalizing and scaling N export predictions to other regions.

HYDROLOGICAL CONSEQUENCES OF FOREST DISTURBANCE AND RECOVERY

The frequency and intensity of regional disturbances in Canada's forests are increasing (Paterson *et al.*, 1998; Schindler, 1998). There are a range of forest disturbances that can alter hydrological processes, including fire and various harvesting practices, involving cut-blocks, patches, shelterwood, or strips. Unfortunately, a basic understanding of the potential impacts of forest disturbance on hydrological processes and their recovery is lacking.

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Potential impacts of forest disturbance on hydrological processes

Disruption of hydrological processes results primarily from changes in precipitation inputs, interception, infiltration and evaporation. Harvesting alters hydrological dynamics during periods of snow cover by affecting radiative fluxes along with snow accumulation, sublimation, snowmelt, soil moisture, soil temperature and soil frost (Meng *et al*., 1995; Pomeroy and Granger, 1997). Comparative snow surveys have shown increases of 30–45% in seasonal snow accumulation after removal of coniferous forest cover at sites across Canada (Pomeroy and Gray, 1995; Pomeroy *et al*., 1997). The effect of forest removal on interception and sublimation of snow is significant and specific for each forest region. For example, Pomeroy and Granger (1997) found that cut-blocks in forests in the semi-arid Boreal Shield ecozone accumulated more snow than in the humid Atlantic Maritime ecozone (Meng *et al*., 1995). Each research group observed an advance in the snowmelt season in disturbed forests resulting from increased net radiation in the cut-blocks. Meng *et al*. (1995) used a forest hydrology model (ForHyM) and a soil temperature model (ForSTeM) to simulate the pre-and post-harvest conditions of a cut-block at the Nashwaak Experimental Watershed Project in New Brunswick. A paired-basin experiment indicated that the cut-blocks produced both positive (as a result of reduced evapotranspiration) and negative contributions (as a result of reduced snow and fog water catch) to the basin water budget. Simulations showed that ground-level insolation advanced the snowmelt season for the cut basin by about two weeks.

Harvesting and its associated activities may also alter the infiltration capacities of the soils. For example, cut-blocks tend to become rutted by machinery or skidding, generally resulting in soil compaction. This reduces soil macroporosity and increases near-surface soil bulk density, making the soil matrix less permeable (Pomeroy *et al*., 1997). Fire may also alter infiltration by promoting hydrophobicity in some soils (Burch *et al*., 1989), thus reducing their infiltration capacities. Conversely, incorporation of logging slash and surface vegetation into the surface layer of the undisturbed forest may decrease bulk density in the top 15 cm of the soil profile (Walley *et al*., 1996). Compaction problems can also be minimized by harvesting during the winter when the upper soil may be frozen and is covered by snow. This practice is often followed in the boreal forest, where the ground surface may be too wet to permit harvesting in other seasons.

Although cut-blocks may have reduced infiltration capacities, they may still store substantial amounts of soil water. At the PAMF, the cut-block site consistently had greater stored soil water than the other regenerating or mature forested sites, perhaps reflecting smaller interception and E fluxes relative to the forested sites. Soil-water storage at regenerating sites may continue to exceed that at mature forest stands owing to the length of time required for growing vegetation to establish an extensive root system and thus increase E fluxes. Much of the excess soil water in disturbed sites is stored below the rooting zone of immature species in cleared and regenerating areas. This water cannot be drawn upon for transpiration and therefore primary production (Pomeroy *et al*., 1997). However, this restriction may not occur for all tree species, as aspen and balsam poplar can regenerate from the existing root system. Harvesting and fire both reduce evaporative losses relative to mature forest canopies. Although such disturbed areas experience greater exposure of the ground surface to wind (Fowler *et al*., 1987) and solar radiation than in mature forests, they have relatively smaller aerodynamic roughnesses and do not experience the direct evaporation of intercepted water and transpiration losses that forest canopies do. Pomeroy *et al*. (1997) observed that burned and cut-block areas differ in their partitioning of energy and water. Burned areas retain some standing timber (thus increasing their aerodynamic roughness relative to cut areas) while not experiencing the soil disturbance that cut areas undergo. Differences in E losses between burned and harvested areas were found to vary with weather patterns and season.

Forested peatlands

Peatlands cover a large portion of Canada, and removal of trees from their surfaces illustrates important changes at the local and regional hydrological scales. Peatlands are organic soils, with saturated and poorly aerated conditions resulting in terrestrial communities that are dominated by tree species such as black

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spruce. Removal of black spruce from peatlands results in reduced interception, evaporation and transpiration, leading to a rise in the water table. Dubé *et al.* (1995) confirmed that cut-blocks in a forested wetland in the St Lawrence Lowlands caused a water table rise, attributed largely to reduced interception. Even intermediate cuttings, such as thinning, resulted in a measurable water table rise. Water table rise is related directly to the percentage of wood cut, depth of the original water table, and time that water is available near the soil surface (Roy *et al.*, 1997). Harvesting of forested wetlands, particularly during the frost-free season, may also disrupt surface drainage, leading to elevated water tables. Groot (1998) examined the physical impacts of site disturbance on peatlands on a black spruce–speckled alder forest, and found that reduced thickness of the aerobic layer and increased water levels (2 to 6 cm) occurred only following intense disturbance.

Greater water availability following tree removal from forested wetlands has considerable consequences for forest recovery. Water table rise may delay regeneration and impair site productivity by reducing the depth of aerated soil needed for tree root activities (Roy *et al.*, 1997). Conversely, forest drainage may create more favourable soil aeration and soil nutrient conditions (e.g. Prevost *et al.*, 1997) and consequently improve tree productivity. Drainage experiments in Canada have been conducted for more than 20 years, and demonstrate significant increases in growth rates. However, despite an increase in the rate of forest drainage from 4000 ha/year in 1985 to 5000 ha/year today (Parent, 1996), the area operationally drained is still small (Prevost *et al.*, 1999). Unfortunately, controlling water table levels through drainage offers only a partial solution at some sites, and emphasis must be placed on avoiding water table increases in the first place (Roy *et al.*, 1997).

Although drainage may have positive effects on tree productivity, it also can lead to subsidence, or a reduction in the surface elevation of peatlands (Rothwell *et al.*, 1996). Subsidence results from the physical breakdown and consolidation of dry peat in the surface layers and accelerated mineralization of the organic matter. The accompanying increase in bulk density is closely associated with the hydrological properties of peat soils (Rothwell *et al.*, 1996). Therefore, subsidence may modify soil hydrological properties beyond the intended effects of drainage alone, with most subsidence occurring within 10 years of drainage. Silins and Rothwell (1998) examined the effects of drainage and subsequent subsidence on the soil hydrological properties of a forested wetland in Alberta. Bulk density and soil-water retention 7 years after drainage were greater than for peat from a nearby undrained control area, associated with loss of macropores (>600 μm diameter) and a concurrent increase in the number of micropores (3–30 μm diameter). Mean saturated hydraulic conductivity was smaller and unsaturated hydraulic conductivity was greater in peat from drained compared with undrained areas. Although these changes in soil-water content and the water transport characteristics resulting from post-drainage subsidence were probably beneficial to the productivity in the specific drained area, considerable variation in post-drainage subsidence will occur among different forested peatland types and climatic regions (Silins and Rothwell, 1998).

Alterations in the drainage of peatlands also may have an impact on streamflow and water quality. Prevost *et al.*, (1999) evaluated drainage effects on water quantity and quality from a back spruce stand in Quebec. Drainage produced an increase in baseflow and the soil's available nutrient content, and had an undetermined effect on peak flow (as a result, in part, of a lack of peak flow events). Suspended sediment concentrations intermittently exceeded acceptable limits only during ditching and precipitation events in the weeks immediately following ditching. Although nutrient concentrations remained within acceptable limits, significant leaching of nutrients occurred during ditching, and concentrations remained above pre-drainage levels 5 years after ditching (Prevost *et al.*, 1999).

Basin studies

Several major whole-ecosystem experiments investigating the potential impacts of forest disturbance on surface waters were initiated during the 1995–98 period, particularly in the Boreal Shield and Boreal Plains ecozones (Figure 10, Table I). Common themes among these experiments were:

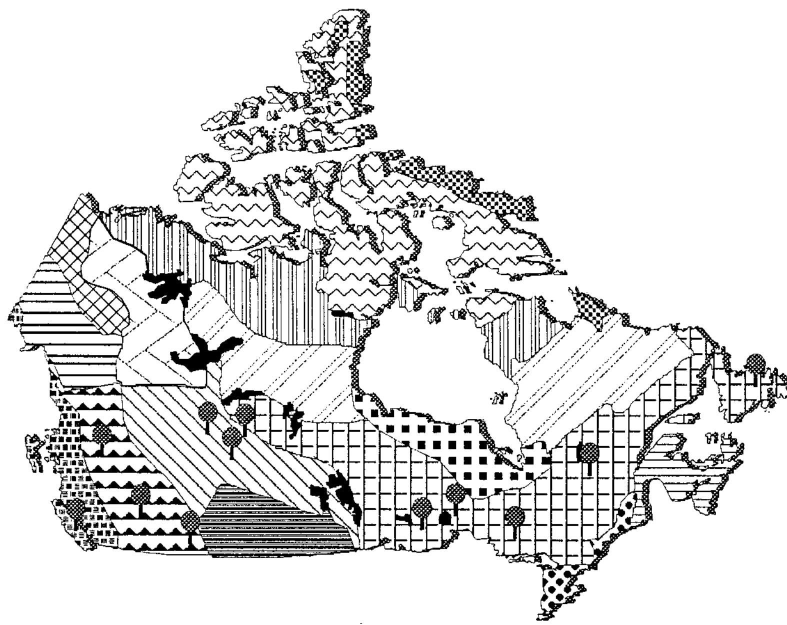

Figure 10. Recent whole-ecosystem studies on the hydrological consequences of forest disturbance and recovery

1. to determine the potential impacts of alternative forest management strategies on stream or lake water quality using a 'paired system' approach (i.e. a control and an experimental system);
2. to compare these potential impacts with natural variability;
3. to extrapolate the findings to a regional scale, where forest management policies are implemented.

Although all studies recognized the influence of changes in the forested landscape on the streams and lakes, the emphasis (and therefore the design) was not on the hydrological linkages between the land and surface waters (e.g. Figure 9). This is not surprising, as evidenced by Swanson's (1998) comment that although 'forest hydrology should be a mature science with routine use of hydrological procedures to evaluate the effect of past, current, and proposed harvesting practices on water resources, . . . it is not' (p. 755). Furthermore, with the exception of the Carnation Creek (British Columbia) and Turkey Lakes Watershed (Ontario) experiments, conclusions are commonly based on data collected for relatively short periods (i.e. 2 to 3 years) during the pre-harvesting and post-harvesting periods (Figure 11). Consequently, these studies will provide only a limited ability to assess the potential impacts of disturbance on basin hydrological dynamics against the backdrop of natural hydrological variation, and to predict the consequences of such changes in hydrological dynamics for water quantity and quality in the basin.

Basin-to-region scaling

Forest management policies intended to protect surface waters are assumed to be relevant over an entire region. At the same time, research results from one or more 'experimental' basins have historically been assumed to be representative of the hydrology and hydrochemistry of the encompassing region. However, recent studies (e.g. Creed and Band, 1998a) demonstrate that there may be substantial sources of natural variation that make the simple extrapolation of results from a single basin to an entire region a questionable exercise.

To scale findings from the basin to the region effectively, the complex interaction of factors (both abiotic and biotic) that influence hydrological dynamics must be considered. For example, interactions between

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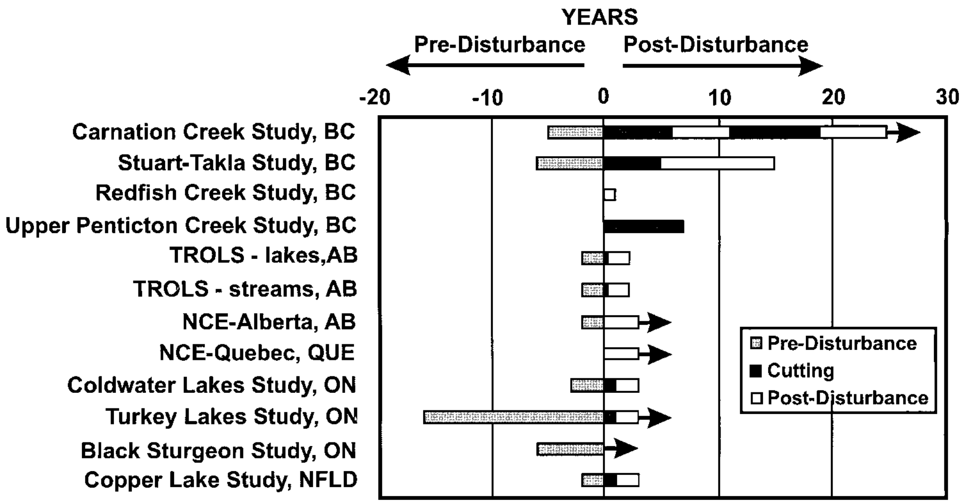

Figure 11. Duration of watershed studies, including pre-disturbance and post-disturbance phases. Post-disturbance phases that do not have an identified termination date are indicated by an arrow (→)

surface and subsurface water systems on the Boreal Plains ecozone are far more complex than those encountered in the Boreal Shield ecozone (Figure 9). Devito and Creed (personal communication) have suggested a hierarchy of landscape factors in this ecozone that regulate the potential susceptibility of surface waters, specifically lakes, to disturbance (Figure 12). The hierarchy consists of factors ranging from coarse to fine scales, including factors that may vary both spatially and temporally within a region (e.g. lake order, elevation, recharge versus discharge nature of the lake, runoff-generation mechanisms from upland and lowland areas to lakes, organization of nutrient formation and flushing areas and shoreline configuration). A lake's position within the landscape may be a major determinant of its potential susceptibility to disturbance (Devito and Creed, personal communication). For example, a lake with a local flow system will be isolated from the effects of a regional flow system, and thus more sensitive to changes within its basin. In contrast, a lake connected to a regional flow system will be less sensitive to changes within its basin, because water inputs from the local area of disturbance will be subsumed by regional water inputs. Landscape hierarchy models of this type can be used as a unifying framework for predicting the potential susceptibility of surface waters to land-cover change over a range of regions. They can also assist in assessing the relative effectiveness of mitigation measures (e.g. riparian buffers).

In addition to landscape factors, the contribution of climatic conditions to natural variations in hydrological dynamics must be considered. Removal of over 90% of the forest cover or complete burns in lake basins in northwestern Ontario led to minor changes in species composition based on palaeolimnological evidence (Paterson *et al*., 1998). Conversely, a gradual change in species assemblages of both reference and disturbed lakes suggested that climate change may have exerted a dominant influence on the lakes (Paterson *et al*., 1998). Westmacott and Burn (1997) evaluated the possible effects of climate change on the magnitude and timing of hydrological events within the Churchill–Nelson River Basin. They found that the magnitude of hydrological events decreased over time with increased air temperature, with the exception of increased flows during spring snowmelt. The need to consider climatic variation when assessing forest disturbance effects on hydrological conditions at the regional scale was stressed by Buttle *et al.* (1998) for the Moose River basin of northeastern Ontario. Thus, changes in peak flows and water yield resulting from harvesting of 10–20% of large (*c.* 10 000 km^2) basins over a 10-year period could not be distinguished from those due to climatic variability with any degree of confidence.

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Table I. Characteristics of some forest ecosystem studies of the potential impacts of forest disturbance indicated in Figure 10

	Carnation Creek, BC	Stuart-Takla, BC	Redfish Creek, BC	Upper Penticton Creek, BC	TROLS, AB
Location	49°N and 125°W	55°N and 125°W	49°N and 117°W	49°N and 119°W	55°N and 113°W
Ecozone	Pacific Maritime	Montane Cordillera	Montane Cordillera	Montane Cordillera	Boreal Plains
Ecoregion	Western Vancouver Island	Fraser Basin	Columbia Mountains and Highlands	Thompson Okanagan Plateau	Mid-boreal Uplands
Temperature	4·2 °C(Jan) to 14·6 °C(Aug)	−10·6 °C(Jan) to 14·2 °C(July)	−3·2 °C(Jan) to 19·9 °C(Aug)	−2·0 °C(Jan) to 20·3 °C(Aug)	−14·6 °C(Jan) to 15·7 °C(July)
Mean annual precipitation	3295 mm	456 mm	732 mm	309 mm	496 mm
Dominant forest species	Western hemlock, western red cedar, amabilis fir, Douglas fir, Sitka spruce, red alder	Sub-boreal spruce, Engelmann spruce	Western larch, Douglas fir, ponderosa pine	Lodgepole pine, Englemann spruce, subalpine fir	Jack pine, white spruce, black spruce, balsam fir, trembling aspen, balsam poplar, white birch
Project design	Paired-basins	Paired-basins	Simulation modelling	Paired-basins	Paired-basins
Project details	Pre-, during and post-harvest comparisons. Treatments included prescribed burning, reforestation, scarification, herbicide use, variable buffer strips and channel disturbance	Pre-, during and post-harvest comparisons. Treatments included variable buffer strips	Modelling of a patch-cut harvest. Scenario 1: 1/3 of basin harvested Scenario 2: 2/3 of basin harvested	Pre-, during and post-harvest comparisons	Pre-, during and post-harvest comparisons. Lakes component: 4 lakes in each of 3 different regions; reference lake, lakes with 20 m, 100 m and 200 m buffer strips Streams component: 5 3rd order streams; each stream with two harvested reaches (upstream, 30 m buffer strip; downstream, no buffer strip)
Project schedule	1971–75: pre-harvest 1976–81: active harvesting 1982–86: post-harvest 1986–93: monitoring and some harvesting 1993–present: monitoring	1990–96: pre-harvest 1996–2001: harvesting	1998–start 1999–end	1995–96: harvesting 1997: monitoring 1998–99: harvesting 2000–03: monitoring 2004–05: harvesting	1994–96: pre-harvest 1996–97: harvesting 1997–99: monitoring

[a] The original experimental design for the Black Sturgeon site had to be modified due to a fire in the area in 1999

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NCE–WAG, AB	Coldwater Lakes, ON	Black Sturgeon, ON	Turkey Lakes, ON	NCE–EAG, QC	Copper Lake, NF
56°N and 113°W	49°N and 92°W	49°N and 88°W	47°N and 84°W	49°N and 75°W	57°N and 58°W
Boreal Shield	Boreal Shield	Boreal Shield	Boreal Shield	Boreal Shield	Boreal Shield
Wabasca Lowland and Mid-boreal Uplands	Lake Nipigon	Lake Nipigon	Lake Temiscaming Lowland	Riviere Rupert Plateau	Southwestern Newfoundland
−19·8 °C(Jan) to 16·6 °C(July)	−17·6 °C(Jan) to 19·2 °C(July)	−21·4 °C(Jan) to 16·0 °C(July)	−10·5 °C(Jan) to 17·8 °C(July)	−19·8 °C(Jan) to 16·6 °C(July)	−7·1 °C(Feb) to 16·2 °C(Aug)
465 mm	685 mm	725 mm	1307 mm	465 mm	1272 mm
Jack pine, white spruce, black spruce, balsam fir, trembling aspen, balsam poplar, white birch	Jack pine, black spruce	White spruce, black spruce, balsam fir, trembling aspen, white birch	Sugar maple, Yellow birch	Jack pine, white spruce, black spruce, balsam fir, trembling aspen, balsam poplar, white birch	Balsam fir
Survey	Paired-basins	Paired-basins	Paired-basins	Survey	Paired-basins
N/A	Pre-, during and post-harvest comparisons. 4 study lakes; a reference lake, 2 lakes with cut blocks to lake edge, 1 lake with 30–50 m buffer strip	Pre-,during and post-harvest comparisons.[a] 4 1st order basins: a reference basin, 1 harvested by conventional clear-cut in summer, 1 harvested by narrow strip partial cut (50% removal) in summer, 1 harvested by narrow strip partial cut (50% removal) in winter	Pre-, during and post-harvest comparisons. 4 1st order basins: a reference basin, 1 harvested by clear-cutting, 1 harvested by selection cutting, 1 harvested by shelter wood cutting	N/A	5 headwater tributaries: pre-, during and post-harvest comparisons 5 headwater tributaries: for 3 streams, treatments included (1) no buffer strip (2) 20 m buffer strip (3) 100 m buffer strip and (4) no harvesting (control reaches)
N/A	1990–96: pre-harvest 1996: harvesting 1997–99: monitoring	1994–00: pre-harvest 2000–01: harvesting	1981–97: pre-harvest 1997: harvesting 1998–present: monitoring	N/A	1993–94: pre-harvest 1995: harvesting 1996–97: post-treatment

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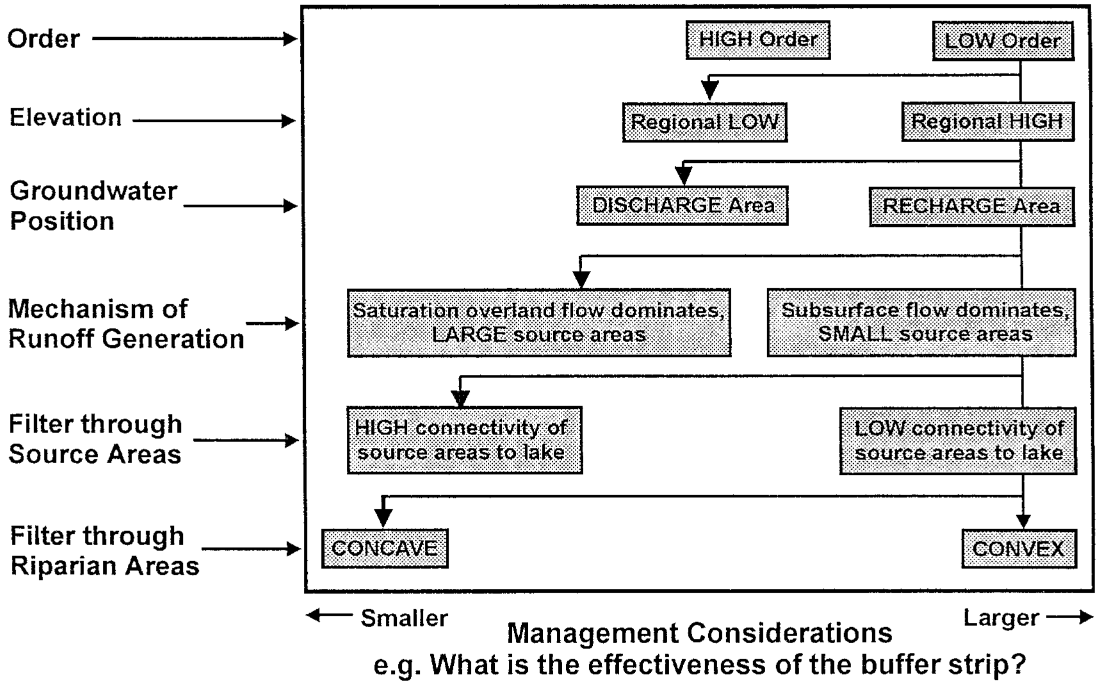

Figure 12. Conceptualization of the hierarchy of landscape factors that influence the potential susceptibility of surface water to land-cover and/or land-use changes (after Devito and Creed, personal communication)

Predicting potential impacts of disturbance with simulation models

A challenge for hydrologists is to develop methods to scale process information and uncertainty, both spatially (from one of several basins to landscape or regional scales) and temporally (from days or years to decades). Simulation models offer a promising tool to predict potential impacts of disturbance (both harvesting and fire) on hydrological dynamics. These include both lumped and distributed models, operating at daily to annual or greater time periods. For example, WRNSFMF (Swanson, 1998), a specific version of the hydrology portion of US Forest Service WRENSS model (US EPA, 1980), is a lumped model that operates at an annual time step and simulates the hydrological consequences of disturbance and recovery. Work at the University of New Brunswick includes model development for forest hydrology (ForHyM; Meng *et al*., 1995), soil temperature (ForStEM; Meng *et al*., 1995), and soil, vegetation and atmosphere dynamics (ForSVA; Arp and Oja, 1997; Oja and Arp, 1997). These lumped models are designed to operate at a monthly to annual time step and have been used to estimate disturbance resulting from cut-blocks (Meng *et al*., 1995) and atmospheric deposition (Oja and Arp, 1997). The RHESSys (Regional HydroEcological Simulation System) is a distributed model incorporating a spatial representation of nested basin and lake systems in a GIS, along with a set of process submodels to compute flux and storage of energy, water, carbon and nutrients (Band *et al*., 1996). RHESSys has been used to explore hydroecological processes and their regulation of basin discharge (Creed *et al*., 1996; Creed and Band, 1998b), and the sensitivity of these hydroecological processes to climate change (Band *et al*., 1996). The model was expanded by coupling carbon and nitrogen dynamics with distributed hydrology to enable prediction of the disruption and recovery of hydroecological processes to time-scales of over 100 years (Mackay and Band, 1997).

Such models provide tools to develop and test hypotheses of hydrological processes in forested landscapes and to predict the potential impacts and interactions of changes in forest characteristics on hydrological dynamics; however, they have been relatively unexploited in research and management applications. A

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primary reason why models are under utilized (or used improperly) is that model parameters, including climate and hydrological data, are rarely available for a particular basin that is to be harvested. Resources need to be committed either to collect these data directly or, perhaps more reasonably, to estimate these data indirectly (Swanson, 1998).

SUMMARY AND CONCLUSIONS

Research into various aspects of the hydrological cycle has been conducted in different forest ecozones across Canada during the 1995–1998 period. Subjects that have received particular attention include the significance of snow interception and sublimation to the water balance of forest basins, the control on infiltration into frozen forest soils, the role of overburden thickness in the hydrological and hydrochemical coupling between forested slopes and receiving wetlands, streams and lakes, and the impacts of harvesting and drainage on water dynamics in forested wetlands. The next several years should see extensive publication of results from this ongoing work, particularly from basin-scale studies investigating the potential impacts of forest disturbance on surface waters.

It is instructive to examine the degree to which the future trends in forest hydrology in Canada suggested in Hetherington's (1987) review have been reflected in recent research activity. These predicted trends include the following.

1. *Greater focus on fisheries–forest interactions, primarily in relation to water quality and aquatic habitat*. As the present review indicates, considerable research has been devoted to the study of the effects of forest disturbance on water quality. In addition, the Eastern and Western Aquatic Groups of the National Centre of Excellence in Sustainable Forestry are currently examining the impacts of forest disturbance on aquatic ecosystems in eastern and western Canada (Table I). However, a limitation of the latter studies from a forest hydrology perspective is that recent advances in our scientific understanding of hydrological processes have generally not been considered sufficiently in studies of the response of basin hydrology to forest disturbance and recovery. This may arise from both the lack of resources to monitor hydrological dynamics properly, and from the failure to involve trained hydrologists in the design of experiments to assess hydrological response to forest disturbance. The key role that hydrologists play in understanding the functioning of forest ecosystems and their response to disturbance and recovery must continue to be stressed to the various levels of government, the forest industry and terrestrial or aquatic-based scientists from other disciplines that are involved the study and management of Canada's forests.
2. *Increased input of hydrological information to management decisions*. Hydrological factors were identified as among the key indicators of sustainable forest management in Canada (Canadian Council of Forest Ministers, 1997). However, Swanson (1998) contends that 'few forest managers give more than lip service to hydrological concerns' (p. 756), and the results of hydrological research have been incorporated into forest management and planning to varying degrees. Hetherington (1987) identified the upsurge in the management of wetlands for forestry as a key example of an area requiring sound scientific information on which to base management decisions. This trend has been borne out by several studies of the hydrological impacts of forestry operations in wetland areas documented in this review, the results of which have been incorporated into peatland management recommendations in northeastern Ontario (e.g. Berry *et al*., 1995). Management implications of studies on the hydrological and hydrochemical consequences of forest disturbance in Canada and elsewhere have also led to recommendations regarding forestry operations in black spruce stands in Ontario (e.g. Nicolson, 1995). In British Columbia, the Clayoquot Sound Scientific Panel's review of all available science (including hydrology) led to recommendations for the use of more environmentally sensitive harvesting systems in the Clayoquot Sound area of Vancouver Island (Anon., 1993). However, the Clayoquot Sound land use decision was hindered by the deadlock that resulted from disagreements over the extent to which forest companies could be expected to follow such recommendations and the fate of undisturbed forested basins in the region (Anon., 1993). Future land use decisions may

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benefit from the experience gained in such efforts as the revision of the Tongass National Forest Plan in southeastern Alaska (Mills *et al*., 1998), where scientists and managers faced a number of critical challenges and learned a number of valuable lessons regarding forest planning and management.
3. *More attention to the renewal and rehabilitation of inadequately reforested land*. We found little progress on this front during the 1995–1998 period, one exception being Buttle's (1995) study of the hydrological and geomorphological effects of headwater reforestation of the Ganaraska River in southern Ontario.
4. *Research into the effects of management activities on forested land*. There is ongoing research in Canada into the hydrological consequences of the extensive road networks that accompany management, including peak flow estimates for the design of forest road bridges and culverts (e.g. Tolland *et al*., 1998). Research in the USA has also examined such issues as the impacts of forest roads on modification on subsurface flow pathways (Beschta, 1998) and basin peak flows (Jones and Grant, 1996), and spatial optimization modelling of the cumulative effects of forest treatments on basin stormflow (Bevers *et al*., 1996). However, we could not find Canadian equivalents of this work.

It is likely over the next decade that forest hydrology research will continue to examine the themes identified by Hetherington (1987). In addition, we suggest that increased attention be devoted to the following areas.

1. *Hydrological processes in a greater range of forested landscapes than have previously been examined in Canada*. A major focus of recent hydrological work has been on boreal forest ecosystems (the Boreal Plains and Boreal Shield ecozones). However, these studies have been conducted largely in the southern portion of the boreal forest, and the relevance of the results for those areas of the boreal forest underlain by permafrost, as well as for the scientifically based management of Canada's other forest ecozones, is unclear. As Black (1998) notes, research into hydrological processes provides the essential context for examination of the implications of forest management decisions.
2. *Dynamics of storm precipitation in mountainous forested landscapes*. This has been identified as a major research priority in the Pacific Northwest of the USA (Beschta, 1998), and is of equivalent relevance to the modelling of floods from forested basins in southwestern British Columbia (e.g. Loukas and Quick, 1996b).
3. *The contribution of climatic change to natural variations in hydrological dynamics in forest ecosystems*. Efforts to model the hydrological consequences of forest management must be made in the context of variations arising from climatic change. Additional impetus for greater research in this area comes from the prospect that the consequences of climate change for the hydrology and hydrochemistry of forested basins may outweigh those arising from disturbances such as harvesting and fire (Paterson *et al*., 1998; Schindler, 1998). Such work depends on long-term interdisciplinary monitoring of the ecological hydrology of forested basins, the importance of which has been recognized in such countries as the USA (Post *et al*., 1998) and Germany (Brechtel and Führer, 1994). However, there are relatively few ongoing monitoring efforts in Canada (Figure 11), and Canadian hydrologists need to stress the scientific and management benefits of sustained hydroecological monitoring.
4. *Development of techniques for transferring data and research results between point, small-basin, landscape and regional scales*. Blöschl and Sivapalan (1995) identified scaling as a major issue in hydrological modelling, and note that it encompasses upscaling (transferring information from a finer to coarser scale), downscaling (transferring information to a finer scale) and regionalization (transferring information from one location to another). Recent work of relevance to forest hydrology in Canada includes the evaluation of alternative means of interpolating climatic data in order to provide spatially distributed data for input to simulation models (Nalder and Wein, 1998), and tests of the geographical transposability of simulation models to other basins in the same region (Donnelly-Makowecki and Moore, 1999). Fractal geometry permits quantification of the relationship of variability between differing scales, such that under certain conditions it is possible to extrapolate the variability of a property or parameter to coarser or finer scales (Blöschl and Sivapalan, 1995). Pomeroy and Gray's (1995) use of fractal analysis in their

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study of sublimation of intercepted snow is a promising example of the applicability of the approach in forest hydrology.

5. *Integration of detailed process studies of the type described in this paper with both conceptual 'landscape hierarchy' models and more detailed simulation models such as WRENSS and RHESSys*. This will improve our insights into the fundamental processes driving the hydrology and hydrochemistry of forested landscapes in Canada. It will also supply the data needed to run and refine such models, which in turn can be used to predict potential impacts of natural and/or anthropogenic disturbances on forest ecosystems. This integration requires us to acknowledge and resolve scaling issues related to hydrological information at the process, measurement and modelling scales, and will depend greatly on progress in (4) above. We must continue to ask whether our scales of measurement or modelling are hydrologically relevant. We also need to collect spatially and temporally intense hydrological information to challenge and, if necessary, modify our conceptual understanding of hydrological processes represented in our models. The generation of such information in turn will hinge on innovations in the instrumentation and technology available to study hydrological processes in Canada's forests.

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