Canadian National Committee for the
International Association of Hydrological Sciences
(CNC-IAHS)

Quadrennial Report to the International Union of Geodesy and Geophysics and International Association of Hydrological Sciences

June 2003

Compiled by

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INTRODUCTION

CNC-IAHS is a committee of the Hydrological Section of the Canadian Geophysical Union (CGU-HS). The main roles of CNC-IAHS are to

1) encourage and promote the participation of Canadian scientists in IAHS,
2) further the collaboration between IAHS and Canadian scientific organizations and institutions, and
3) respond, on behalf of Canada, to scientific requests of IAHS.

CNC-IAHS also has an administrative role in seeking and supporting the nominations of Canadian hydrologists to executive positions of IAHS and arranging the selection and nomination of National Representatives and National Correspondents to IAHS Commissions and Committees. IAHS is one of the seven associations of the International Union of Geodesy and Geophysics (IUGG).

The CNC-IAHS Executive

The Executive of CNC/IAHS consists of the Senior and Junior Canadian National Representatives (NR) for IAHS plus the President and Vice-President of CGU-HS, the Presidents of the Canadian Meteorological and Oceanographic Society (CMOS), the Canadian Water Resources Association (CWRA) and the Canadian Chapter of the International Association of Hydrogeologists (CCIAH), and one Member-at-Large elected from the general membership of CGU-HS. The Senior NR serves as the Chair of the Committee and the Junior NR as Secretary. The current Chair of CNC-IAHS is Professor Taha Ouarda, Institut national de la recherche scientifique (INRS-ETE), Université du Québec; his term of Office will terminate in April 2005. The Secretary is Professor John Pomeroy, Dept. of Geography, University of Saskatchewan, Saskatoon; he will act as Junior NR until 2005 when he will replace Professor Ouarda as Senior NR.

Major CNC-IAHS activities

Presently, the main activities of CNC-IAHS concern the XXIII General Assembly of IUGG at Sapporo, Japan, in July 2003. These activities are the contribution of CNC-IAHS to the quadrennial report of the Canadian National Committee for IUGG (CNC-IUGG) to be tabled at Sapporo, and the Elections of IAHS Officers and those of the IAHS Commissions to be held during the Plenary Administrative Session of IAHS.

The *IAHS election* procedure was initiated in the fall of 2002 when CGU-HS and CNC-IAHS circulated a call for the nomination of candidates to stand for office in the IAHS Bureau and that of the IAHS Commissions. All terms of office are for the period 2003-2007 except for position of President, which involves serving two years as President-Elect (2003-2005) and four years as President (2005-2009). The result of the call for nomination resulted in no candidates for the IAHS Bureau elections but two candidates were nominated for office in two IAHS Commissions. D. de Boer was nominated for the position of Secretary of the International Commission on Continental Erosion (ICCE) and A. Pietroniro for the position of President of the International Commission on

Remote Sensing (ICRS). In addition two other Canadian candidates were nominated through a separate IAHS procedure. These are: J. Barker for the position of Vice-President of the International Commission on Groundwater (ICGW) and J. Gibson for the position of President of the International Commission on Tracers (ICT). As all these four Canadian candidates are unopposed for Office they will be elected by default.

Other current activities

A *Canadian Journal of Hydrology*: CNC-IAHS has opened discussions with CGU-HS and CWRA on the opportunity of publishing a Canadian journal devoted to the hydrological sciences. Publication of research in Canadian hydrology is presently fragmented between many journals of varying scientific quality. However, the excellent quality of papers by Canadian hydrologists is generally accepted internationally and would guarantee a high scientific credibility and widespread circulation of such a publication.

There is interest amongst Canadian hydrologists in the IAHS Decade for Prediction of Ungauged Basins as a substantial portion of Canada is poorly or ungauged compared to more densely populated countries. Professor John Pomeroy of the University of Saskatchewan was appointed to the Science Steering Group of PUB and has been asked by the International Commission on Snow and Ice to promote cold regions hydrology issues within PUB.

The *Montreal AGU-CGU Meeting, 2004*: CNC-IAHS and CGU-HS are looking closely at the need of attracting good Canadian convenors and proposals for special sessions on both Canadian hydrology and the more universal scientific problems of water resources. There is interest in having a special PUB session in Montreal.

Structure of the CNC-IAHS Quadrennial Report to IUGG

The key elements of this report consist of a series of review papers on the progress of Canadian hydrology for the period 1999-2003. Certain members of the Canadian hydrological community were solicited by CNC-IAHS to prepare the works. The papers and authors are:

1) Advances in Canadian Forest Hydrology 1999-2003 (J.M. Buttle, I.F. Creed, R.D. Moore),
2) Advances in Canadian Wetland Hydrology (J.S. Price, B.A. Branfireun, J.M. Waddington, K.J. Devito),
3) Snow, Frozen Soils and Permafrost Hydrology in Canada (M.K. Woo and P. Marsh),
4) Advances in River Ice Hydrology (B. Morse and F. Hicks),
5) A Revised Canadian Perspective: Progress in Glacier Hydrology (D.S. Munro),
6) Recent Canadian Research on Contemporary Processes of River Erosion and Sedimentation (D. de Boer, M. Hassan, B. McVicar, and M. Stone),
7) Progress in Isotope Tracer Hydrology in Canada (J. Gibson, T.W.D. Edwards, S.J. Birks, N.A. St Armour, B. Buhay, P. McEachern, B.B. Wolfe and D.L. Peters).
8) A Review of Canadian Remote Sensing and Hydrology, 1999-2003 (A. Pietroniro and R. Leconte)

These same eight papers will also be submitted for peer review and publication as an insert in the CGU-HS Special Issue of *Hydrological Processes* to be published in late 2003 or early 2004. The Guest Editorial Board for this issue will be S. Beltaos, L. Martz, D. Moore, and T. Ouarda.

Submitted by H.G. Jones and J.W. Pomeroy for the Canadian National Committee-IAHS.

Advances in Canadian Forest Hydrology, 1999 – 2003

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Abstract:
Understanding the hydrological processes and properties of Canada's varied forest types is critical to sustaining their ecological, economic, social and cultural roles. This review examines recent progress in studying the hydrology of Canada's forest landscapes. Work in some areas, such as snow interception, accumulation and melt under forest cover, has led to modelling tools that can be readily applied for operational purposes. Our understanding in other areas, such as the link between runoff-generating processes in different forest landscapes and hydrochemical fluxes to receiving waters, is much more tentative. The 1999 – 2003 period saw considerable research activity examining the hydrological and biogeochemical response to natural and anthropogenic disturbance of forest landscapes, spurred by major funding initiatives at the provincial and federal levels. This work has provided valuable insight; however, application of the findings beyond the experimental site is often restricted by such issues as a limited consideration of the background variability of hydrological systems, incomplete appreciation of hydrological aspects at the experiment planning stage, and experimental design problems that often bedevil studies of basin response to disturbance. Overcoming these constraints will require, among other things, continued support for long-term hydroecological monitoring programs, the embedding of process measurement and modelling studies within these programs, and greater responsiveness to the vagaries of policy directions related to Canada's forest resources. Progress in these and related areas will contribute greatly to the development of hydrological indicators of sustainable forest management in Canada.

INTRODUCTION

Forest covers 417.6 million ha of Canada's land area (almost half the country), and plays critical ecological, economic, social and cultural roles at the local, regional, national and global scales. Sustaining these roles depends in part on knowledge of the dominant hydrological processes and properties of the forest landscapes in Canada's various ecoregions. This knowledge is critical to understanding such diverse issues as forest productivity; the quantity and quality of water moving to receiving wetlands, streams and lakes; the role of forests in surface – atmosphere exchanges of energy, water and carbon; and the physical and biogeochemical implications of disturbance to forest landscapes. This last theme is of particular relevance given the scale and intensity of forest disturbance in Canada. Thus, ~0.4% of Canada's forests is harvested each year, while ~0.5% is disturbed by fire or insect outbreaks (Natural Resources Canada 1998). There is evidence that the balance between these forms of forest disturbance is changing. For example, Schroeder and Perera (2002) noted that while the area burned within the managed forest area of Ontario remained relatively constant at 0.5 million ha decade^{-1} between 1951 and 1990, the total clearcut area increased from 0.5 million ha (1951-1960) to >2 million ha (1981-1990). Our ability to understand, predict and manage the varied consequences of these and other natural and anthropogenic disturbances to forest ecosystems, as well as our efforts to sustain the ecological, economic, social and cultural roles of Canada's forests, need to be based on sound hydrological principles.

This review builds on an earlier examination of the state of forest hydrology in Canada (Buttle et al. 2000), and identifies progress on many of the major themes discussed in that review. Much of this progress has been driven by several important funding programs initiated shortly before or during the period covered by the present review. These have had an important influence on the location, intensity and objectives of research related to hydrological properties and processes in Canada's forest landscapes. Examples at the provincial level include the Forest Renewal British Columbia (FRBC) program, which began in the early 1990s and provided a major impetus to forest hydrology research in that province, and the Ontario Living Legacy program, which was initiated in the late 1990s and has supported several projects related to the hydrological consequences of forest disturbance. The major initiative at the federal level was the institution of the Sustainable Forest Management Network (SFMN). Its mission is to develop networks of researchers from universities, industry, government and First Nations that will promote sustainable resource management strategies for Canada's forests. This network has supported considerable research into the hydroecological implications of forest disturbance in Canada, the initial results of which were summarized in a special issue of the *Canadian Journal of Fisheries and Aquatic Sciences* (Carignan and Steedman 2000).

PRECIPITATION AND SNOW PROCESSES

Rainfall Interception, Throughfall and Stemflow

The limited Canadian research between 1999 and 2003 on interception, throughfall and stemflow during rainfall in forest stands was mainly concerned with the role of these processes in forest nutrient cycling in forests and the effects of acid deposition on nutrient fluxes from the canopy to the forest floor. The nature of specific hydrologic processes received relatively little attention; nevertheless, the work is still relevant to forest hydrology. Thus, Houle et al. (1999a) assessed the number of collectors required to measure throughfall depth associated with a predetermined error and confidence level as a function of the measurement time interval for a mixed hardwood stand in Quebec. The results assist in designing sampling strategies for estimating throughfall inputs in forest landscapes. Houle et al. (1999b) examined ion deposition in precipitation, throughfall and stemflow in deciduous and coniferous stands in the same drainage basin. Average annual interception in the deciduous and coniferous stands was 11 and 18% of total precipitation, respectively, while stemflow contributed 3 and 1% of the annual net precipitation reaching the forest floor in the deciduous and coniferous stands, respectively. All values agree with those reported in the literature. Gordon et al. (2000) examined throughfall and stemflow fluxes in equal-age plantations of red, black and white spruce in central Ontario. Absolute water fluxes were not reported; however, the relative ranking of throughfall fluxes was black spruce > red spruce > white spruce, while the relative ranking of stemflow fluxes was white spruce > red spruce >> black spruce. These differences were attributed to the unique morphologies (e.g. bark roughness, branch angle, crown structure) associated with each species, and reinforce a recent call (Levia and Frost 2003) to consider morphological properties when assessing interception, throughfall and stemflow studies conducting in differing forest types.

Yanni et al. (2000)'s use of the ForHyM2 model to simulate throughfall and streamflow in four forested basins in south-western Nova Scotia provides an exception to the research focus of the previous studies. Three of the basins in the Yanni et al. (2000) work were dominated by spruce and pine with variable amounts of hemlock and fir, while the fourth basin was dominated by maple, oak, birch and beech. Modelled throughfall agreed well with measurements from nearby basins (Percy 1989), and modelled average annual interception was 9.4% of mean annual gross precipitation. It is worth noting that the authors had to add fog drip contributions of between ~150 to ~180 mm year^{-1} to water inputs to the basins to obtain good agreement between observed and simulated streamflows. More empirical studies are needed to develop and validate models of this important but often-overlooked process in maritime forests on Canada's east and west coasts.

Snowfall, Snow Interception, and Snowmelt

Recent progress on snow accumulation and melt in forest landscapes can be placed into three somewhat-overlapping categories: process-based research; integrated process- and modelling-based research; and modelling- and monitoring-based research for operational purposes.

Process-Based Research

Winkler (2001) studied snow accumulation and melt in pine and spruce-fir mature stands, juvenile stands and clearcuts at Mayson Lake and Upper Penticton Creek, south-central British Columbia. Smaller peak snow water equivalent (*SWE*) occurred in forest cover than in clearcuts, with the exception of a juvenile spruce-fir stand. Average snowmelt rates in forested stands were 0.4 – 0.9 × those in the clearcut at Mayson Lake and 0.6 – 0.7 × those in the clearcut at Upper Penticton Creek, with no difference in melt rate between juvenile spruce-fir stands and the clearcut. Daily melt rates from continuous lysimeter measurements showed earlier and more rapid melt in the juvenile-thinned pine stand relative to the juvenile-unthinned pine stand and clearcut, while clearcut melt rates exceeded those in all other stands later in the season. Stand structural properties provided significant explanations of standardized ratios of forest-to-clearcut peak *SWE* and melt. A radiation budget model that incorporated the standardized ratio of forest-to-clearcut melt successfully predicted measured snowmelt at Mayson Lake.

Murray and Buttle (2003) noted that most research into forest harvesting impacts on snow accumulation and melt has been for coniferous forest stands. They compared snow accumulation and melt in adjacent hardwood maple and clearcut stands in central Ontario, focusing on the role of slope aspect and canopy density in controlling inter-stand differences in melt rates. As expected, accumulation and melt in the clearcut exceeded that in the forest stand; however, the control of aspect on between-stand differences in melt rate was far greater than that of canopy.

Faria et al. (2000) built on previous work suggesting that melt rates within forest stands decrease with increasing *SWE* (e.g. Buttle and McDonnell 1987) and that *SWE* varies at both the stand and intra-stand scales. They measured changes in the spatial distributions of *SWE* before and during melt in five stands of differing canopy density in central Saskatchewan. A log-normal distribution fit the pre-melt frequency distribution of *SWE* within stands. Greater variability in *SWE* resulted in earlier exposure of ground under spatially-uniform melt simulations; however, the spatial distribution of daily melt was inversely correlated with *SWE*, which further accelerated snow cover depletion. Inclusion of within-stand covariance of melt and *SWE* improved estimates of measured changes in snow-covered area relative to simulations that only considered the effect of *SWE* distribution on snow-cover depletion.

Integrated Process- and Modelling-Based Research

The distinction between this category and the next is that the modeling work in these studies is largely intended to complement our understanding of hydrologic process. Woo and Giesbrecht (2000a) used energy balance estimates of snowmelt in a subarctic spruce woodland that incorporated differential melt rates due to canopy shading. These estimates were then used to examine the relationship between variability in mean daily snow depth and the spatial scale at which snow depth was determined, and assessed the information loss occurring with increasing spatial scale of aggregation. Information loss increased with snow depth variability as melt progressed, and this scale-induced information loss should be considered when modelling snowmelt in regions exhibiting large spatial variations in melt rates. Woo and Giesbrecht (2000b) presented a model to simulate snowmelt under a subarctic spruce tree, based on physical melt processes, canopy geometry, and field-derived empirical functions and coefficients. Simulations compared well with measured daily snow depths, and the tree canopy enhanced the snow surface's longwave radiation balance. Changing melt intensities produced by the tree's presence induced a strong asymmetry in melt rates among difference azimuths within and beyond the tree canopy. A companion paper (Giesbrecht and Woo 2000) used a simplified version of the model to simulate melt in a subarctic spruce woodland using GIS. The model included topographic and tree shadow effects on the radiative component of melt energy. Meteorological data from an open site were used to estimate melt for zone types in the forest. Skewness of the snow depth distribution decreased as melt progressed, while variability in snow depth increased. The authors cautioned that the error associated with using point values of snowmelt to estimate spatially-averaged melt exceeded 200%.

Parviainen and Pomeroy (2000) extended the research of Pomeroy and colleagues into snow interception and sublimation in forest landscapes summarized in the previous progress report on Canadian forest hydrology (Buttle et al. 2000). They coupled physically-based equations describing snow interception and sublimation processes with a one-dimensional land surface scheme (the Canadian Land Surface Scheme CLASS – Verseghy 1991, Verseghy et al. 1993). The coupled model was tested against measured sublimation in mature and regenerating jack pine stands in central Saskatchewan. It provided good simulations for the mature stand, but did not estimate latent heat fluxes well during events involving larger snow loads and incoming solar radiation. This was attributed to errors introduced by solving for within-canopy humidity and to the role of subcanopy snow energetics not considered in the coupled model.

Modelling- and Monitoring-Based Research for Operational Purposes

The ForHyM2 model (Arp and Yin 1992) is an operational hydrology model designed to simulate water fluxes in forest ecosystems using limited input data. Bhatti et al. (2000) used it to simulate snowpack depth in a jack pine site in northeastern Ontario. The model gave good predictions of measured peak *SWE*, with greater *SWE* in open areas relative to beneath the canopy, as observed elsewhere (see Murray and Buttle 2003 for review). Similar agreement between measured and modelled *SWE* occurred when ForHyM2 was applied to forested basins in south-western Nova Scotia (Yanni et al. 2000). Pomeroy et al. (2002) developed a model for predicting forest snow accumulation based on stand properties (canopy density, leaf area index) and seasonal snow accumulation or seasonal snowfall in nearby small clearings (S_c). The work was done in the Wolf Creek basin in the Yukon and in central Saskatchewan. The model was based on physically-based snow interception equations derived from previous work by Pomeroy and colleagues, and was consistent with Kuz'min's (1960) relationship between snow accumulation in a forest (S_f) and canopy density (C_c):

$$S_f = S_c(1-0.37C_c) \qquad [1]$$

The authors suggested that relationships of this form are spatially transferable between cold climate forests.

Hudson (2000) examined aspects of hydrologic recovery during stand regeneration. This is an important theme in forest hydrology that is closely linked to the issue of sustainable forest management. Hudson compared snow accumulation and melt in regenerating stands in coastal British Columbia across a range of canopy heights, based on the assumed relationship between canopy height and degree of stand regeneration. Recovery factors were calculated using linear interpolations between extremes defined by the peak accumulation or melt rate of old growth and clear-cut equivalent plots. There was rapid initial recovery, and an asymptotic exponential model provided a reasonable description of recovery as a function of either canopy height or canopy density. The results suggested a hydrologic recovery threshold where the height of the tallest trees in the stand is roughly equal to the mean peak snowpack depth for open sites. Relationships of this type can be used to estimate the degree to which regenerating stands have reached complete hydrologic recovery. However, tree height provided relatively poor predictions of hydrologic recovery in juvenile forest stands in south-central British Columbia (Winkler 2001). The largest proportion of variability in forest peak *SWE* relative to that in the open was explained by crown volume, length and closure, while melt was best predicted by the square root of basal area. Winkler suggests that incorporation of stand-structure variables that represent snow interception and shading under varying forest cover conditions provides a sounder means of estimating hydrological recovery than use of tree height alone.

The H_{60} concept has been used in western Canada and the United States to relate snow-covered area in mountain basins to basin peak flows during spring melt. The H_{60} line is the elevation above which 60% of the basin lies, and Garstka et al. (1958) found that peak stream flow occurrence from a Colorado basin corresponded to a snow-free area of roughly 60%. Thus, the snow-covered area above the 60% line can be considered to be the major source area of water contributing to peak flows. The H_{60} concept has been used to guide the planning of forest harvesting operations in the southern British Columbia interior, based on the untested assumption that forest removal above the H_{60} line will increase *SWE* and melt, and thus runoff contributions to peak flows. Gluns (2001) evaluated the applicability of the H_{60} concept to conditions in interior British Columbia using measurements of snowline elevation and streamflow for five snowmelts in two small basins near Nelson, British Columbia. About 65% of each basin was snow-covered at the time of peak flow, suggesting that the H_{60} concept is a valid planning tool for evaluating forest-harvesting plans in south-central British Columbia.

RUNOFF PROCESSES AND BASIN WATER BALANCE

There does not appear to have been as much research on this topic as was summarized in the progress report. Despite this, many of the same issues related to runoff production and streamflow generation highlighted by Buttle et al. (2000), such as the role of various runoff processes in forest landscapes, the degree of hydrological coupling between hillslopes and receiving waters, and the influence of this coupling on basin streamflow characteristics, have also been examined in several studies since 1998. Much of this work has formed part of larger multi-disciplinary studies, such as the Boreal Ecosystem – Atmosphere Study (BOREAS) and the Canadian GEWEX programme.

Forested Precambrian Shield

A major impetus for research into runoff processes in forested basins continues to be an interest in the response of terrestrial and aquatic ecosystems on the Canadian Shield to acid deposition. This work recognizes the important role that runoff processes and water flowpaths exert on the chemistry of water moving through terrestrial ecosystems and discharging to streams, lakes and wetlands. Much of this research during the past five years was conducted at two long-term research sites: the Turkey Lakes Watershed in central Ontario, and the Muskoka-Haliburton area of south-central Ontario.

Hazlett et al. (2001) related the chemical composition of stream and soil water from two first-order basins in the Turkey Lakes Watershed during snowmelt to assess flowpaths to streams. Snowmelt inputs bypassed the deeper soil zone in the high-elevation basin, reducing buffering of snowmelt acidity. There was relatively greater input of deeper soil water in the low-elevation basin, resulting in higher pH and base cation concentrations and smaller Al levels in streamflow. Semkin et al. (2002) used silica as a conservative tracer in a mixing model approach to estimate contributions to streamflow from pre-melt streamflow, water routed through the forest floor, and water routed through the upper mineral soil during five snowmelts in the high-elevation basin studied by Hazlett et al. (2001). On average, pre-melt streamflow, and water routed through the forest floor and upper mineral soil contributed 9, 28, and 63%, respectively, of basin discharge. A greater proportion of forest floor water was delivered at maximum stream discharge. Semkin et al. (2002) argued for the development of a perched water table above the contact between the surficial ablation till and underlying basal till in the basin. This shallow groundwater initially delivered water to stream through the upper mineral soil, and eventually (at maximum discharge) rose to intersect the ground surface and deliver runoff as return flow. Flow from the deeper till was insignificant, consistent with soil water and streamflow chemistry relationships presented for this basin by Hazlett et al. (2001). Buttle et al. (2001b) examined groundwater characteristics during spring snowmelt for the two basins studied by Hazlett et al. (2001). Isotopic signatures were used to estimate groundwater residence times. Depth to the piezometric surface was shallower and there was a marked increase in groundwater residence time with depth in the high-elevation basin, consistent with the shallow flowpaths suggested by Hazlett et al. (2001) and Semkin et al. (2002). Greater depths to piezometric surfaces and similar residence times for shallow and deeper groundwater in the low-elevation basin also agreed with Hazlett et al.'s (2001) suggestion that deeper water flowpaths play a greater role in this basin. This study also found no consistent relationships between groundwater characteristics and the Beven and Kirkby (1979) $\ln(a/\tan\beta)$ topographic index for either basin.

Studies of runoff processes in the Muskoka-Haliburton area of south-central Ontario have focused on the role of preferential pathways in transporting water through forest slopes. Buttle and Turcotte (1999) examined the control of throughfall and pre-event soil water characteristics on throughfall partitioning between overland and subsurface flow and between bypassing flow and translatory flow at the slope scale. Overland flow occurred under intense throughfall and decreased antecedent soil wetness, suggesting this pathway was most effective during drought conditions which promoted hydrophobicity of the organic layer. This finding echoes that of Biron et al. (1999), who hypothesized that soil hydrophobicity limits infiltration in forest soils during dry conditions in order to account for rapid surface runoff (as inferred from streamwater chemistry) over unsaturated near-stream soils in a small forested basin near Montreal, Quebec. Buttle and Turcotte (1999) noted that vertical bypassing flow was independent of pre-event soil water content, but was directly related to throughfall intensity. The strong association between throughfall intensity and slope runoff suggested that coupled vertical and lateral macropore flow controlled runoff generation during small-to-medium size events. Their hypothesized increase in translatory flow displacement of pre-event soil water on the slope with larger events and greater antecedent wetness was partly supported by Buttle et al.'s (2001a) summary of natural and artificial tracer studies on the slope. Comparison of the spatial distribution of soil macroporosity on the same slope with measured and modelled point infiltration characteristics showed that sites with greater surface-derived macroporosities had increased bypass flow to depth (Buttle and McDonald 2000). Examination of the links between this vertical preferential flow and slope scale runoff during artificial irrigations showed that lateral macropores made a minor contribution to slope runoff, which was dominated by flow in a thin layer above the soil-bedrock interface (Buttle and McDonald 2002). This flow occurred in a highly conductive zone at the bedrock surface and in the overlying soil matrix, and showed complex mixing of event and pre-event water. Vertical macropore flow was independent of antecedent soil wetness (consistent with Buttle and Turcotte 1999); however, antecedent soil wetness combined with soil depth and bedrock topography to determine the thickness, connectivity and upslope extent of the pre-event saturated layer above the bedrock surface. These properties of the saturated layer in turn controlled whether vertical preferential and matrix flow reaching the bedrock surface participated in slope runoff.

Hypermaritime Forest Landscapes

The benefits of combined hydrometric and tracing approaches to examine hydrological processes were highlighted in Gibson et al.'s (2000) study of runoff generation on bog – forest uplands in the Prince Rupert area of coastal northern British Columbia. A three-component isotope hydrograph separation showed that shallow hillslope groundwater accounted for 85% of peak streamflow during a mid-summer rainfall. Analysis of baseflow discharge and isotopic response also permitted estimation of groundwater mean residence time and soil storage capacity in the study basin. An important contribution was the use of systematic shifts in deuterium excess of rainfall for labeling shallow and deep groundwaters based on their residence time signatures. The work also reinforced the value of using reactive tracers (in this case dissolved organic carbon - DOC) to study water cycling processes.

Forested Landscapes Underlain by Permafrost

Research has focused on the role of permafrost in controlling temporal and spatial patterns of slope runoff and its contribution to streamflow generation. Carey and Woo (1999) examined north- and south-facing subarctic forested slopes in the Wolf Creek basin in the southern Yukon. The south-facing slope had earlier snowmelt but no lateral surface or subsurface flow, since meltwater infiltrated the seasonally frozen soil cover with low ice content. Summer moisture exchanges were dominated by infiltration and evaporation. Conversely, deep percolation during snowmelt was hindered on the north-facing slope, which had an organic layer overlying clay sediments with permafrost. This led to surface runoff in rills and gullies and subsurface flow through pipes and the organic soil matrix. Soil pipes occurred at the organic and mineral soil horizon interface, and transmitted water only when the water table was within or above this interface (Carey and Woo 2000). Pipe flow was described by the Manning equation combined with estimates of contributing areas, which were relatively small but changed in extent with slope wetness. Pipeflow made a significant contribution to slope runoff during snowmelt; however, matrix flow within the organic layer dominated runoff when ground thaw lowered water tables. Carey and Woo (2000) concluded that models of hydrologic processes in permafrost environments must consider the control the frost table exerts on the phreatic surface, and thus on the potential for pipeflow. Subsurface flows on the north-facing slope during the summer were confined to the conductive organic layer. Slope evaporation decreased as both frost and water tables descended during the summer. Initial results of Carey and Woo (1999) were extended (Carey and Woo 2001) through examination of slope water balances during snowmelt and summer periods on four contrasting aspects. Snowmelt runoff was confined to slopes with organic soils with an ice-rich base that prevents meltwater infiltration. Flow was unimpeded through frozen but porous organic materials, and lateral runoff was initiated when the organic layer's storage capacity was exceeded. Summer runoff was confined to wet organic-covered slopes, with larger flow from slopes where the water table was close to the surface. For some slope segments, inflow from upslope relative to outflow at the slope base significantly affected the slope water balance. The slope remained wet and runoff was enhanced when inflow equaled or exceeded outflow; however, minor inflow from upslope led to lowering of the water table, reduced runoff in the organic layer and drying of near-surface soils. Permafrost slopes and organic horizons were the principal controls on streamflow generation in subarctic basins.

Interactions between *SWE*, rainfall magnitude and timing, thaw depth and antecedent levels in surface water stores also controlled water balance dynamics and streamflow generation in a boreal forest basin in northern Manitoba (Metcalfe and Buttle 1999, 2001). Inter-annual differences in the amount and timing of rainfall inputs relative to active layer thickness largely dictated the degree to which surface stores on slopes and in wetlands become filled by initial meltwater inputs. These differences in storage capacity in turn controlled whether subsequent meltwater and rainfall were exported as streamflow. Isotopic and geochemical hydrograph separations showed that meltwater dominated streamflow during conditions of intense melt on slopes with limited soil thawing combined with large pre-melt storage in surface depressions (Metcalfe and Buttle 2001). Conversely, smaller melt intensities combined with deeper active layers and smaller storage levels in basin wetlands led to subdued streamflows largely supplied by older water routed through less-permeable deeper peat layers and mineral soil.

Methodological Issues Related to Runoff Processes in Forest Landscapes

Hydrological research in forest landscapes increasingly requires data on soil water content. Time domain reflectometry (TDR) is becoming the method of choice for obtaining such data, due to its ease of operation and ability to be coupled with data loggers for continuous measurement of water content at various depths and locations. Spittlehouse (2000) reviewed the methodological challenges facing the use of TDR to measure soil water content in stony soil, and discussed various factors to be considered when calibrating the TDR. Greater soil water contents were measured in a clearcut relative to a forest site in southern interior British Columbia, consistent with previous work suggesting that removal of forest canopy should increase soil water contents due to reduced interception and evaporation losses (e.g. Elliott et al. 1998). Instead, Spittlehouse attributed the clearcut soil water content to the smaller average stone content of its soils, and highlighted the need to consider soil properties when interpreting the hydrological consequences of forest disturbance.

Estimation of runoff fluxes and water balance components in Canada's forest regions is severely limited by a shortage of hydrometric monitoring stations, as well as the short record length often associated with such stations (Prowse 1990). These stations are also frequently located on large rivers (>1000 km^2), and hydrologic data are generally unavailable for lakes and streams that may be impacted by harvesting operations at the scale of a forest management plan (10 – 100 km^2). In response to this data limitation, Gibson (2001) described the use of the evaporative enrichment of stable environmental isotopes (oxygen-18 and deuterium) in surface waters as an indicator of water balance variations in forest and tundra landscapes of northern Canada. He reviewed the assumptions underlying the approach, along with the applicability of an isotopic steady-state model to lakes of varying size. The method provided reasonable estimates of basin-wide evaporation rates, and permitted discrimination between evaporative and transpirative losses when combined with total evapotranspiration estimates from hydrometric methods. This approach was extended by Gibson et al. (2002) to estimate throughflow, residence time and basin runoff to 70 headwater lakes in northern and north-central Alberta. Runoff to lakes in wetland-dominated basins exceeded that in upland-dominated lakes, with generally greater runoff from basins with low bog/fen ratios. However, there was no clear indication that forest disturbance (either by harvesting or fire) resulted in a significant change in runoff/precipitation ratios relative to reference lake basins. Nevertheless, Gibson et al. (2002) contended that the approach provides a useful added tool for studying the hydrological controls on lake chemistry and ecology, and for assessing the hydrological consequences of forest disturbance.

There is increasing use of hydroecological models to simulate runoff production and streamflow generation in forest landscapes, both in Canada (e.g. Alila and Beckers 2001, Whitaker et al. 2003) and in other countries (e.g. Tague and Band 2001). Such models often use digital topographic data to simulate spatial variations in basin wetness and to route water from slopes to the stream channel. However, the degree to which surface topography reflects the hydraulic gradients driving shallow subsurface flow in forest basins is often unclear. Hutchinson and Moore (2000) examined this issue based on measurements of hillslope outflow at nine throughflow troughs installed at a road cut on a forested slope in the lower mainland of British Columbia. They also measured the spatial distribution of the water table draining to the troughs. The upslope contributing area derived from topography of the underlying basal till provided a reasonable description of throughflow distribution across the slope at low flows; however, surface topography was a better approximation of water table form at high flows. Estimates of effective hydraulic conductivities (K_H) at slope widths <10 m varied over two orders-of-magnitude and showed no consistent relationship with saturated layer thickness. Conversely, there was a linear increase in K_H with saturated layer thickness for greater flows at slope widths of ~10 m. All K_H versus saturated layer thickness profiles contradicted the parabolic and power-law transmissivity profiles sometimes assumed by hydrological modelers. Hutchinson and Moore (2000) also noted that shunting of water by discrete macropores can overwhelm topographic controls on throughflow at slope widths <10 m.

FOREST HYDROCHEMISTRY

Our knowledge of the impacts of acid deposition and climate change on biogeochemical cycling in forests is based on a limited number of basin studies. Significant variability in sulphur (S) and nitrogen (N) export among basins within a relatively small region has been reported within eastern Canada (e.g., Creed and Band 1998a, 1998b, Devito et al. 1999, Beall et al. 2001, Watmough and Dillon 2002) and therefore information obtained from a single basin study may not accurately reflect the sensitivity of the majority of forest basins in a given region to acid deposition and/or climate change (Watmough and Dillon 2002). Wetlands, a common feature in Canadian forests, represent critical interfaces between slopes and receiving streams and lakes. Significant relationships have been observed between the proportion of wetlands in basins and export of DOC (Prepas et al. 2001a, Creed et al. 2003), phosphorus (P) (Devito et al. 2000, Evans et al. 2000, Prepas et al. 2001a), N (Prepas et al. 2001a), and S (Devito et al. 1999). The reader is directed to Advances in Canadian Wetland Hydrology, 1999-2003 (Price et al. this issue) for a more comprehensive discussion of biogeochemical cycling in and export from wetlands. Studies focused on the interactions of hydrological and biogeochemical cycles and their impacts on basin export of S and N in forests are discussed below.

Sulphur

Forests have received reduced S deposition since the early 1980s, leading to the expectation of reductions in sulphate (SO_4) export from basins; however, SO_4 export downstream of wetlands in some basins persists (Dillon and LaZerte 1992). Mass balance studies showed a net export of SO_4 following summer droughts in a forested swamp located in a basin with shallow tills (< 1m), but net retention of SO_4 in a forested swamp in a basin with deeper tills (≥ 1 m) (Devito 1995). Devito and Hill (1999) examined the relationship between water table elevation, SO_4 mobilization versus immobilization in the surface peat of wetlands, and SO_4 export from basins in central Ontario. They also estimated the depth to which SO_4 mobilization occurs in the wetland in order to quantify SO_4 pools available for future export pulses. They found a critical threshold response of increased SO_4 mobilization when the water table declined such that the capillary fringe no longer extended to the surface (i.e., ≥ 25 cm), resulting in drainage and aeration of the peat. The total S pool in wetlands suggested that recovery periods resulting from recent reductions in S deposition by streams draining wetland-dominated basins may be significantly longer than those observed for upland-dominated basins (Devito and Hill 1999).

In order to predict which landscapes are most susceptible to prolonged acidification from atmospheric deposition, Devito et al. (1999) explored how widespread the SO_4 pulse phenomenon was and which wetlands were susceptible to water table drawdowns and therefore episodic SO_4 release following droughts. They found that classifying basins based on < or ≥ 50% coverage of uplands with a till depth > 1m helped to identify which basins with wetlands produce large SO_4 export following dry conditions. Basins with predominantly shallow tills (< 1 m) in the uplands were characterized by transient upland-wetland hydrologic connections that promoted large water table drawdowns, re-oxidation of accumulated S in the wetlands, and large SO_4 export. In contrast, basins with predominantly deep tills (≥ 1m) were characterized by continuous upland-wetland hydrologic connections, leading to smaller water table drawdowns and smaller SO_4 export (Devito et al. 1999). Eimers and Dillon (2002) expanded this conceptual model by examining a gradient of basins varying from no wetland coverage to significant wetland areas and till depths ranging from < 1 m to ≥ 1 m. The basins showed a high degree of synchrony in inter-annual patterns of SO_4 export, suggesting that processes affecting the entire basin (i.e., both upland and wetlands) are involved in net SO_4 export. They suggest that: (1) reduced atmospheric S loading in basins with no wetlands may result in desorption of SO_4 that was previously adsorbed when S loading was higher as the soils shift toward an altered "equilibrium state" (*sensu* Reuss and Johnson 1986); and (2) warmer and drier climatic conditions in basins with or without wetlands result in higher mineralization rates of organic S compounds in both upland and wetland soils and greater S export (Houle et al. 2001, Eimers and Dillon 2002).

Nitrogen

In contrast to S, N deposition rates have not changed since the 1980s, leading to continued concerns regarding N saturation of forest ecosystems (Aber et al. 1989). Increased nitrate-N (NO_3-N) export from forest slopes to receiving surface waters is a diagnostic for N saturation of forests (Stoddard 1994), but the *source* of NO_3-N and the *processes* by which NO_3-N is mobilized from the slope to the stream remain unclear.

Creed et al. (2002) characterized the spatial heterogeneity in total N and potentially mineralizable N (PMN) pools in soils in a deciduous forest in central Ontario. They hypothesized that topography regulates the spatial pattern of these pools through a combination of static factors (slope, aspect and elevation) that influence radiation, temperature and moisture conditions, and dynamic factors (catenary position, profile and planar curvature) that influence downslope transport of materials. This hypothesis was tested using statistical models to explore the topographic basis for the pattern of N and PMN pools. Random sampling resulted in multiple linear regression and tree regression models that produced similar totals (i.e., within 5% of each other) but dissimilar patterns of the pools, with the latter producing a more realistic heterogeneous distribution of N. Static factors were the most important predictors of the pattern of N pools; however, the authors acknowledge that a hydrologically-based sampling strategy may have produced a model where both static and dynamic factors were predictors.

Spoelstra et al. (2000) used $^{15}N/^{14}N$ and $^{18}O/^{16}O$ isotopic ratios of nitrate-N to determine the source of NO_3-N exported from deciduous basins in central Ontario. Although external N (i.e., atmospheric deposition of NO_3-N and ammonium-N [NH_4-N]) was significant, internal N (i.e., NO_3-N produced by nitrification of NH_4-N) was the dominant source of NO_3-N exported from basins. Lamontage et al. (2000b) used NO_3-N labeled with ^{15}N to evaluate where N inputs are stored in a coniferous basin in western Ontario. In the short term (few years), the surface organic horizons of forest soils were the main sinks for N inputs to the forest; however, a comparison of a N mass balance analysis with ^{15}N recovery revealed a significant missing sink for ^{15}N which the authors suggested was fine or coarse woody debris on the forest floor. Lamontagne and Schiff (1999) demonstrated considerable spatial heterogeneity in N sinks within a forest landscape. Forest "islands" were N sinks while the lichen, moss, and grass community on surrounding bedrock outcrops were N sources. Although forest islands covered a small proportion of the basin, they had a major impact on NO_3-N export because most of the water leaving the basin had to move through at least one forest island before leaving the system (Lamontagne and Schiff 1999). Future studies must consider the pattern of N pools, N sources *versus* N sinks, and the hydrological connectivity among these functional groups for more effective models of N export from basins.

The N flushing hypothesis (Creed et al. 1996) has been used to explain the natural variability in N export from basins. N flushing may be regulated by matrix processes *via* water table fluctuations or by macropore processes *via* preferential flow pathways. Mechanisms of mobilization from N pools to streams were investigated by Hill et al. (1999) and Buttle et al. (2001a). Hill et al. (1999) hypothesized that NO_3-N flushing occurs by macropore preferential flow pathways that mobilize NO_3-N rapidly from the surface soil horizon both down the soil profile and down the slope, and that the N chemistry of subsurface stormflow was controlled by mixing of event water moving *via* macropore flow with pre-event soil matrix water. There was no evidence of NO_3-N flushing, and NO_3-N concentrations in subsurface storm flow were small despite input inorganic N concentrations several orders of magnitude larger than in the mineral soil water. As water infiltrated the soil, high rates of microbial immobilization of NO_3-N, low rates of net N mineralization, and no net nitrification in the surface soil horizon resulted in small inorganic N concentrations in pre-event soil water and therefore little or no NO_3-N flushing. The N in subsurface storm flow at the soil-bedrock interface consisted of N in *event* water transported via macropores (Hill et al. 1999; Buttle et al. 2001a). These studies emphasize the need to consider how hydrology, topography (of both the surface and bedrock surface) on forest slopes interact with soil processes to control N export.

These complex interactions between acid deposition impacts in the face of climatic variability and extreme climatic events in the short term, and climate change in the long term, mean that generalization of observed short-term patterns to longer time scales must be approached with caution (e.g., Biron et al. 1999, Courchesne et al. 2001).

HYDROLOGICAL AND HYDROCHEMICAL ASPECTS OF FOREST DISTURBANCE

Several studies during the past four years have improved our understanding of surface water (stream and lake) response to basin disturbance, including deforestation by clearcut harvesting and wildfire (cf. Carignan and Steedman 2000). Most of this work has focused on the boreal forest, which comprises 32% of Canada's forest cover (Natural Resources Canada 1998). Current strategies for forest management assume that harvesting activities emulate wildfire and therefore will sustain boreal forest dynamics. This assumption is based on the effects of wildfires on terrestrial ecosystems, and does not consider aquatic ecosystems (cf. Pinel-Alloul et al. 2002). Several major research programs were initiated to investigate the effects of human activity on the hydrological and biogeochemical linkages between the land and waters in forested landscapes in Canada.

Natural Variability of the Hydrologic System

We examined the 30-year hydrological patterns in different forest regions of Canada to provide a hydrological template to facilitate comparison of the conclusions of published studies. First, we analysed total annual precipitation (P), potential evapotranspiration (PET), and discharge (Q) to assess the dominance of these components of the hydrologic budget in different forest regions. Patterns of P, PET, and Q suggest significant differences in the magnitude of land-atmospheric versus land-aquatic hydrologic exchanges among the forest regions. For example, selected sites in the boreal forest show $P<PET$ in the Boreal Cordillera, $P \cong PET$ in the Boreal Plain, and $P>PET$ in the Boreal Shield. Changes in the hydrologic balance as one moves from the western to the eastern edges of the boreal forest will have important implications for predicting the potential impacts of forest disturbance on these hydrological systems.

Second, we examined the cumulative departures from normal monthly precipitation (*CDNP*) and temperature (*CDNT*). Positive slopes of *CDNP* and *CDNT* indicate wetting and warming conditions, respectively, while negative slopes for both graphs indicate drying and cooling climatic conditions compared to the long-term average (Winter et al. 2001). There is substantial heterogeneity in climatic conditions, and therefore our selected sites should not be considered representative of the each physiographic region. We chose sites closest to the geographic region within which previous studies of the impacts of forest disturbance on hydrology were conducted. Cumulative departures from monthly precipitation or temperature are useful in depicting the naturally-occurring oscillations in climatic conditions. For example, the selected stations have precipitation signals that show: (1) large variability with multiple major overlapping climatic oscillations ranging from a few years to decades (e.g., Boreal Shield); (2) moderate variability with a single minor climatic oscillation of about 10 years (e.g., Boreal Plain); and (3) small variability with no apparent climatic oscillations (e.g., Boreal Cordillera). The selected stations show similar variability in the temperature signals. Of interest is the fact that sites in the boreal forest with relatively small variability in precipitation show relatively large changes in temperature (e.g., Boreal Cordillera) and those with relatively large variability in precipitation show relatively small variability in temperature (e.g., Boreal Shield).

Our analysis highlights the importance of discriminating the disturbance "signal" from the naturally variability of background "noise" when quantifying the impacts of forest disturbance on hydrology and the physical, chemical, and biological characteristics of aquatic ecosystems that hydrology regulates in Canadian forests.

Disturbance by Harvesting and Wildfires

There were few studies of the potential impacts of timber harvesting on surface waters to report in the previous review on forest hydrology in Canada (Buttle et al. 2000). In contrast, in this review has noted many studies on this topic during the 1999 – 2003 period, including those from several watershed manipulation experiments. The introduction to a Special Issue of the *Canadian Journal of Fisheries and Aquatic Sciences* on the impacts of forest disturbance on aquatic ecosystems [Volume 57 (Suppl. 2), 2000], identified the main science questions underlying contemporary concerns about the sustainability of aquatic ecosystems as: (1) has human activity compromised the ability of forests to produce clean water and support productive and diverse aquatic biota?; and (2) will new stresses have a synergistic or antagonistic effect on existing stresses? The following summarizes studies that address these two questions.

Response to Disturbance – Streams

Water Quantity. Prevost et al. (1999) used a paired-basin approach to examine the effects of draining a forested peatland in Quebec (Table 1). Baseflow was increased by 25% following ditching in the treatment basin. Peak flows did not appear to be affected by the drainage; however, a definitive conclusion could not be drawn due to the lack of high flow events during the calibration period.

McFarlane (2001) used a retrospective paired-basin approach to find control and treatment basin pairs from 66 gauged basins in southeastern British Columbia. After finally chosing two basin pairs, he found that although specific changes in peak flow could be identified, these changes could not be conclusively related to removal of forest cover. The results did not support a threshold level of harvesting above which changes in peak flows could be detected. McFarlane noted that high statistical power for detecting hydrologic changes requires more years of data collection than are generally available in paired-basin experiments. This can lead to a failure to reject the null hypothesis of no change when in fact change may have occurred.

Caissie et al. (2002) examined streamflow changes at Catamaran Brook, New Brunswick, for two sub-basins subjected to clearcutting of 2.3% (Middle Reach) and 23.4% (Tributary 1) of the basin areas. Annual and seasonal water yield at Middle Reach did not change following harvesting, based on comparisons of post-harvest data with pre-logging regressions, which used water yield for the Little Southwest Miramichi River (basin area 1340 km^2) as the predictor variable (i.e., statistical control). Regressions between peak flow and stormflow volume for Middle Reach against storm rainfall did not differ significantly between pre- and post-logging periods. Comparison of pre- and post-harvest regressions of Tributary 1 peak flow against both storm rainfall and Middle Reach peak flow indicated that harvesting increased peak flow magnitude. However, there was no statistically-detectable effect on storm flow volumes at Tributary 1.

Buttle and Metcalfe (2000) examined streamflow response to boreal forest disturbance (fire and harvesting) in six sub-basins of the Moose River basin, northeastern Ontario, two with "medium" drainage areas (400-1100 km^2) and four with "large" (6800-12000 km^2). They determined the extent of forest disturbance in each basin using remote-sensing images from two dates, and analysed streamflow data using an after-the-fact pairing for the medium and large basins, with the least disturbed basin in each size group serving as a control. Disturbance effects on annual runoff and peak flows could not be detected by double-mass curves and trend analysis. However, changes in small and medium flows in two basins appeared to be related to forest disturbance in the medium and one large basin, respectively.

Whitaker et al. (2003) explored the effects of different forest harvesting scenarios on streamflow by applying the Distributed Hydrology-Soil-Vegetation Model (DHSVM) to Redfish Creek, a mountainous snowmelt-dominated basin in the southern interior of British Columbia. Comparisons of modelled and observed *SWE* and snow-covered area agreed reasonably, supporting the validity of the simulated snowmelt inputs. The simulations revealed significant year-to-year variations in the effect of harvesting on streamflow, particularly peak flow, in agreement with empirical results from 30 years of post-harvest data at Fool Creek, Colorado (Troendle and King 1987). Simulations indicated that harvesting below the H_{60} elevation did not influence peak flows, since the snow had melted from that zone by the time of peak flow. This supports use of the H_{60} elevation as the basis for a weighting factor in the calculation of equivalent clearcut area within the Interior Watershed Assessment Procedure (British Columbia Ministry of Forests and Ministry of the Environment 1999).

The activity reported here represents a significant improvement on research into the effects of forest disturbance on streamflow regimes cited in Buttle et al.'s (2000) previous review. Nevertheless, there is still a relative shortage of empirical studies on disturbance impacts (both natural and anthropogenic) on water yields, peak and low flows in Canada's various forest landscapes. This contrasts with the lively and ongoing debate regarding the influence of forest management on peakflows in the Cascade Mountains of western Oregon (Jones and Grant 1996, Thomas and Megahan 1998, Beschta et al. 2000), and the recent call in the United States to examine harvesting impacts on flooding during extreme events under a range of management practices, physiographic conditions, and event types (DeWalle 2003). The absence of studies specific to the Canadian environment means that inferences about forest disturbance impacts on a basin's hydrologic regime are often drawn from work conducted in other parts of the world. This is illustrated in Scherer's (2001) review of studies on the impacts of forest-cover removal on peak flow magnitude and timing, water yield and low flows relevant to conditions in the central and southern interior of British Columbia. Only five of the 18 studies were conducted in Canada, with the remainder from Arizona, Colorado, Idaho, Montana, Oregon and Utah. The issue of importation of research results to the Canadian context is an important one. In some cases importation is hampered by the absence of analogues to the Canadian situation, such as the large-scale harvesting and fire disturbance occurring in the boreal Shield and boreal plains landscapes. In cases where harvesting under similar forest cover and climatic conditions is being done elsewhere, direct importation of research results may not be appropriate. For example, some hydrologists in British Columbia are questioning the extent to which the non-glaciated Oregon Coast Range provides a good model for glaciated British Columbia environments. In addition, many U.S. studies of forest harvesting impacts on basin streamflow conducted in the 1960s and 1970s employed practices, particularly in relation to road construction and maintenance and site preparation (e.g. slash burning), that are no longer in use.

Stream Temperature. Four studies used basin-scale experiments to document the effects of forest management on stream temperatures (Table 1). Prevost et al. (1999) found that draining a forested peatland decreased weekly minimum temperature by 2°C and increased weekly maximum temperature by 7°C. Water temperature in the drained basin often reached 25°C or more. Bourque and Pomeroy (2001) detected slight warming after harvesting in four streams, despite retention of generous riparian buffers. They attributed the warming to advection of heated subsurface water from the cutblocks. They also related variations in pre- and post-harvest temperatures among treatment streams to indices of solar irradiance on the streams and cutblocks and mean slope gradient in the cutblocks. Mellina et al. (2002) found that both of their study streams cooled in the downstream direction, even after harvesting reduced riparian canopy cover to about half its pre-treatment value. They attributed this cooling to the presence of small lakes upstream of the study reaches, where temperatures became greater than the equilibrium temperature for conditions in the stream reaches. A synoptic survey of lake-headed and non-lake-headed streams having a range of forest management histories confirmed that stream reaches exhibit downstream cooling for some distance below small lakes, even through cutblocks. Despite this downstream cooling trend, there was a net warming of up to about 2-4°C (in terms of daily maximum temperature) during August at the downstream ends of the cutblocks, in comparison with predicted temperatures based on a pre-logging regression with the control. Curry et al. (2002) found that fall season water temperatures were greater in a stream with no buffer, but a treatment effect could not be detected in a stream with a 20-m buffer. To isolate treatment effects, Curry et al. used an analysis of covariance (ANCOVA) on daily mean water temperature, with daily mean air temperature as a covariate. Curry et al. also found that temperatures in brook trout incubation habitats were similar to surface water temperatures, reflecting the dominance of downwelling hyporheic flow over upwelling groundwater.

Several studies modelled stream temperature. Caissie et al. (2001) related maximum daily water temperatures at Catamaran Brook, New Brunswick, to air temperatures using both logistic regression and stochastic models. The stochastic model included a sine component for the seasonal cycle and a second-order Markov process to account for short-term deviations. St-Hilaire et al. (2000), also focusing on Catamaran Brook, modified the CEQUEAU parametric-conceptual model to account for lateral advection of heat associated with subsurface flow, to be able to simulate the effect of groundwater warming in clearcuts on stream temperatures. Inclusion of heat inputs by lateral inflow improved stream temperature simulations. Mitchell (1999) developed a regression model based on monthly data to predict the effect of clearcut harvesting with no buffers.

Suspended Sediment. Kreutzweiser and Capell (2001) studied fine sediment infiltration into the streambed at six locations representing a gradient of forest harvesting impacts in the Turkey Lakes Watershed, Ontario. The greatest sedimentation rates for the inorganic fraction were recorded downstream of road-improvement activities and at a site where skidder tracks in the riparian zone channelled flow into the stream. The lowest sedimentation rates occurred at a shelterwood treatment, where logging roads were not a factor. Harvesting activities did not appear to affect the organic fraction of the particle-size distribution.

Prevost et al. (1999) found that ditching a forested peatland to improve drainage increased suspended sediment concentrations by 100 – 200 × for a few weeks following during ditching. Concentrations then returned to pre-drainage levels, except during one rainstorm in the second year after treatment.

A number of sediment-related studies have been conducted through funding from FRBC. Most of these studies have not been reported in peer-reviewed journals, although some results have been presented in Forest Service Technical Reports or conference proceedings (e.g., Hudson 2001, Henderson and Toews 2001, Jordan 2001). Christie and Fletcher (1999) examined the effect of harvesting activities on sediment geochemistry in five small streams in the Stuart-Takla Fish-Forestry Interaction Project, located in the sub-boreal spruce biogeoclimatic zone of British Columbia. Prior to logging, each stream had distinctive sediment chemistry associated with basin geology. Harvesting alone did not change sediment geochemistry, probably because the presence of riparian buffers reduced sediment inputs to the channels. Road crossings did, however, change sediment geochemistry through the introduction of abraded zinc from galvanized culverts, and by introducing sediment via erosion of the road bed and ditches. Analysis of patterns of sediment geochemistry both along the streams and through time allowed rates of sediment transport and dispersion to be calculated. Although this study focused on potential impacts of logging activity on geochemical exploration, it also identified the potential value of sediment geochemistry as a tracer for forestry-related sediment disturbances.

Methodological Issues. Many of the experimental studies involved relatively short pre-treatment periods (one – two years). Three used regression analysis or ANCOVA using daily time series to control for interannual variations in weather and hydrologic conditions, while Bourque and Pomeroy (2001) "standardized" their data by dividing temperatures in the treatment streams by the corresponding temperature in the control stream. One potential problem is that the daily time series, and the error terms in the regression analyses or ANCOVAs, are likely autocorrelated. This reduces the effective degrees of freedom compared to those calculated from sample size, which would bias the estimated significance levels for hypothesis tests. Future studies should consider using time-series regression approaches (e.g., generalized least squares) for paired-basin calibration using short pre-treatment periods. Another problem is that the pre-treatment regressions may not be stable from year to year.

Buttle and Metcalfe attributed the lack of detectable effects to the buffering of flow changes by the large basins and the relatively low levels of disturbance (maximum of 25% of the drainage area). Other issues are the confounding by inter-basin differences in physiographic characteristics and climatic variability in both space and time, which could lead to differing streamflow trends amongst basins even in the absence of disturbance effects. These issues are unavoidable when studying medium to large basins, where rigorously-controlled experiments are not feasible. However, an important point is that lack of a detectable effect is not the same as no effect. A potential strategy is to supplement statistical analyses with modelling studies. For example, a modelling exercise along the lines of that conducted by Whitaker et al. (2003), if done prior to an experiment or in conjunction with a statistical analysis of historical hydrometric data, may provide at least a first estimate of the likely magnitude and direction of a treatment effect as an aid to interpreting statistical results (e.g., by conducting a statistical power analysis along the lines of McFarlane 2001).

One weakness of the modelling exercise conducted by Whitaker et al. (2003) is that they were only able to test the model's capacity to simulate current conditions at Redfish Creek. As Klemeš (1986) stressed, such a test does not guarantee that a model can accurately predict streamflow under changed conditions (i.e., after logging in this case). Future model applications should focus on paired basin situations, where the treatment effect can be estimated statistically through comparison with a control. Such work is currently being conducted by Alila and co-workers at the University of British Columbia Faculty of Forestry, using data from the Carnation Creek and Penticton Creek experiments in British Columbia.

Response to disturbance - Water quality

In addition to the increases in water levels, water and sediment yield, and water temperatures that often accompany timber harvesting, harvesting may also impact hydrologic linkages between terrestrial and aquatic systems and thus influence ecosystem productivity and integrity. An important indicator of changes to such linkages is provided by the chemical composition of receiving surface waters. Increased particulate nutrient loads are often associated with erosion resulting from water flowing over soil surfaces that have been compacted during harvesting and associated activities (e.g. roads) (Jones and Grant 1996). Increased dissolved nutrient loads are associated with enhanced microbial conversion of nutrients at the soil surface from non-mobile to mobile forms, and subsequent export when the water table rises to the soil surface and flushes mobile nutrients over the saturated soil surface to the stream or lake (Hornberger et al. 1994, Creed et al. 1996). However, changes in the chemical composition of surface waters following harvesting vary substantially between localities, even within a physiographic region.

Comparison of Boreal Shield vs. Boreal Plain. Studies on the effects of forest disturbance on the chemical composition of receiving surface waters have focused on the Boreal Shield and the Boreal Plain (Table 2). Drainage basin characteristics of boreal lakes vary. Lakes on the Boreal Shield are smaller and deeper, with shorter water residence times, and their drainage basins are smaller and steeper with smaller wetland coverage than on the Boreal Plain (Table 3). Similarly, indicators of water quality vary. Lakes on the Boreal Shield have smaller concentrations of DOC, total phosphorus (P), total nitrogen (N), and chlorophyll *a* than those on the Boreal Plain (Table 4). Despite these differences, significant trends in the response of these lakes to forest disturbances have emerged.

The number of indicators of water quality affected by forest disturbance was greater in the oligotrophic Boreal Shield lakes compared to the eutrophic Boreal Plain lakes (Table 5). For example, Boreal Shield lakes had increased loadings of DOC, P, and N following both harvest and wildfire disturbances (Steedman 2000, Carignan et al. 2000a, Lamontagne et al. 2000, Enache and Prairie 2000). In contrast, Boreal Plain lakes had no change in DOC or N loading, but had increased P loading following harvest disturbances (Prepas et al. 2001a) and increased DOC, P and N loading following wildfire disturbances (McEachern et al. 2000). The degree to which the impacts of increased loadings cascaded up the trophic structure was limited, with generally greater phytoplankton biomass, variable zooplankton biomass, and no change in fish biomass (Planas et al. 2000, Patoine et al. 2000, St. Onge and Magnan 2000, Prepas et al. 2001a). There was inter-lake variability within each boreal subregion, with the response of some water quality indicators in disturbed lakes strongly related to the lake's drainage ratio (i.e., ratio of drainage area to lake area), where lakes with the highest drainage ratios showed the largest response (Table 5).

Current strategies for sustainable forest management are based on the assumption that management practices will sustain boreal forest dynamics if they emulate natural disturbance regimes (e.g., wildfire) (Hunter 1993). Consequently, recent studies have focused on comparing the effects of harvesting and wildfire on aquatic ecosystems (e.g., Carignan et al. 2000a; Prepas et al. 2001a; McEachern et al. 2000). It is perhaps not surprising that the intensity of disturbance effect in terms of magnitude and/or duration was greater for wildfire than for harvesting given that the proportion of the basin disturbed by wildfire was greater than for harvesting (Table 5).

Results from empirically-based studies such as those previously described cannot be extrapolated in time or space. Consequently, there is a scientific need for incorporation of process-based monitoring and modeling approaches into the experimental design of studies so that: (1) the "noise" in the chemical response of lakes resulting from climatic variability and/or climate change with each sub-region of the boreal forest can be effectively characterized (e.g., process-based models can be used to simulate hydrological processes across the complete range in climatic conditions); and (2) the "signal" in the lakes' chemical response to forest disturbance can be effectively discriminated from the noise (e.g., models can be used to simulate both the "signal" and the "noise" by using different simulation scenarios).

Effectiveness of Forest Buffer Strips. Riparian forests buffers, strips of forest left between harvested blocks and adjacent surface waters, have been considered important sinks for particulate and dissolved nutrients in hydrologic flows. Therefore they have been used to mitigate changes resulting from increased water and nutrient loads to lakes and streams following timber harvest (Castelle et al. 1994). The widespread acceptance that buffer strips protect aquatic resources is paralleled by the recognition that guidelines regarding buffer strip design have not been established based on scientific merit. Consequently, buffer strips often satisfy neither those who want these areas protected nor those who want access to the timber held in these protected areas (Buttle 2002).

Guidelines requiring forest industries to leave a standard width of buffer strip around surface water bodies are often based on "rule of thumb", and studies that have applied these guidelines to the boreal forest suggest they may not be effective. While one study implied that standard width buffer strips reduce nutrient loadings to lakes (Carignan et al. 2000a), other studies indicated that these buffers do not reduce nutrient loading to lakes (e.g., Steedman 2000, Steedman and Kushneriuk, 2000, Prepas et al. 2001a, Pinel-Alloul et al. 2002). There is a lack of fundamental knowledge on the impacts of timber harvesting on lakes and the effectiveness of buffer strips in mitigating these impacts in the boreal forest.

Buttle (2002) points out that use of a standard width in the application of the forest buffer concept does not consider such basic hydrologic concepts as: (1) lake area relative to the area of the forest draining to the lake (i.e., the drainage ratio), (2) the degree to which harvested areas are hydrologically-connected to the stream or lake; or (3) the relative input from local, intermediate and regional groundwater systems to the stream or lake. He contends that these hydrological considerations can assist in designing effective buffer strips. Recognizing that hydrologic data are often not available for this task, Buttle (2002) offers simple proxies of hydrological conditions that can be generated using automated analysis of digital elevation data. For example, topographic indices (*TI*) (e.g., $TI = \ln [a/\tan\beta]$, where a is the upslope area draining to a given location and β is the slope gradient at that point, Beven and Kirkby 1979) can be used to identify groundwater discharge versus recharge areas along the land-water interface (Buttle 2002). Wolniewicz (2002) combined the mapped *TI* with topographic depressions and flat slopes and used a simple inundation routine to generate a time series of maps of recharge versus discharge areas. These maps can be used as a basis for designing buffer strips that mitigate water, sediment and nutrient loadings to surface waters. Hydrologically-based designs of buffer strips will likely move away from the practice of forested ribbons along streams and rings around lakes to forest patches that protect critical recharge or discharge zones within the landscape. Furthermore, there is a recognition that the design of buffer strips for mitigating water, sediment and nutrient loadings to surface waters may not be sufficient to alleviate other impacts (e.g., increases in wind speeds over the lake which have implications for internal lake processes), and that more research on the role of buffer strips in allaying all ecological impacts that harvesting may pose to receiving waters is needed (Buttle 2002).

Criteria and Indicators for Sustainable Forest Management (SFM). Despite the potential ambiguity of "sustainable forest management", the term has been retained in national and international initiatives and defined as the maintenance of a series of criteria and indicators (Kneeshaw et al. 2000). A criterion is a category or class of processes characterized by a set of related indicators that are monitored periodically to assess change. An indicator is a quantitative or qualitative variable which can be measured or described and which, when observed periodically, demonstrates trends (http://www.mpci.org/criteria_e.html). In Canada, the Canadian Council of Forest Ministers (CCFM) have identified criteria of SFM, including: (1) conservation of biological diversity; (2) ecosystem condition and productivity; (3) conservation of soil and water resources; (4) global ecological cycles; (5) multiple benefits of forests to society; and (6) accepting society's responsibility (CCFM 1997). Major research programs, including the SFMN Centres of Excellence, have focused on defining indicators for each criterion.

Kneeshaw et al. (2000) comment that scientific knowledge must be incorporated into the development of indicators. They define the essential attributes of indicators of SFM as being: (1) scientifically sound; (2) operationally feasible; (3) socially responsible and internationally credible; (4) measured following a standard method; (5) easily measurable and cost-effective; (6) easily interpretable and directly linked to environmental changes generated by *local* management activities but relatively insensitive to more *global* sources of variation; (7) integrated; and (8) linked to prescriptions. For the criterion related to the conservation of water resources, indicators must be identified that preserve the functional and structural integrity of aquatic ecosystems and the natural resources associated with these ecosystems. To show threshold and achieved target levels of indicators, this preservation must incorporate the *natural range of variation* of the functional and structural properties as determined by their response following *natural disturbance regimes* (Kneeshaw et al. 2000).

Table 6 summarizes studies that have focused on developing hydrologically-based indicators to predict potential changes in surface water quality in response to natural and/or anthropogenic disturbance regimes. Many have focused on the drainage ratio, or the ratio of the size of the drainage basin that drains into a lake relative to the size of the lake (area or volume). Large drainage ratio lakes (large basin area: small lake area) have enhanced potential for nutrient loading *via* surface drainage to the lakes and short water residence (or water renewal) times. Conversely, small ratio lakes (small basin area: large lake area) have reduced potential for such nutrient loading and longer water residence times. This ratio was correlated to surface water quality parameters on both the Boreal Shield and Boreal Plain (cf. Pinel-Alloul et al. 2002).

As one moves from comparatively simple hydrological systems on the Boreal Shield to the more complex Boreal Plain, more indicators may be required to represent the hydrological controls acting on a lake (Buttle et al. 2000). Devito et al. (2000) developed a conceptual model of the hierarchy of these controls and identified indicators that represent each level of this hierarchy, including: (1) the hydrogeologic setting of a lake, which defines the relative importance of subsurface flow contributions to the lake from local, intermediate and regional flow systems; (2) the hydrologic efficiency of surface water drainage from the basin's contributing source areas to the lake; and (3) the surface versus subsurface pathways of water moving from source areas to the lake. Recognition of this hierarchy of landscape controls was important in predicting the potential loading of nutrients to lakes on the Boreal Plain (Devito et al. 2000). Subsequent studies recognized that nutrient source areas vary with the hydrologic conditions required for mobilization of specific nutrients. For example, source areas for DOC, dissolved organic N and total P may be characterized by topographic depressions and flat zones (i.e., saturated areas with longer water residence times), while nitrate-N and ammonium-N source areas may be characterized by gentle slopes with large upslope contributing areas (i.e., saturated areas with shorter water residence times) (Creed and Beall 2003). This section has focused on "static" indicators that do not incorporate the natural variability of the hydrological system. However, research is being conducted into developing "dynamic" indicators that relate the natural variability in hydrologic fluxes from forests to surface waters to the natural variability in the contributing source areas within the forest (e.g., Krezek 2000, Wolniewicz 2002).

CONCLUSIONS

We began this review by noting that our ability to understand and manage the response of forest ecosystems to natural and anthropogenic disturbance, and to manage forest resources in a sustainable fashion, will depend on a sound grasp of hydrological principles operating in Canada's varied forest landscapes. We have made some progress in obtaining a better understanding of hydrologic conditions in these landscapes, as this review has indicated. However, further progress will require us to come to terms with the following research issues:

1. We need to improve our understanding of the subsurface hydrology. Several studies cited here have highlighted our inability to anticipate the hydrologic response of a forested slope or basin without considering the nature of its subsurface flowpaths. We must develop tools or techniques for characterizing such flowpaths. A starting point would be determining what (if any) relationship exists between surface features and subsurface flowpaths. We also need to improve our predictive understanding of the nature of fractured bedrock, macropores, and hyporheic zones and their possible role in regulating a basin's hydrologic response. This is particularly important for attempts to evaluate the hydrologic consequences of various forest management practices, since we must be able to predict the within-basin variability in these subsurface flowpaths prior to conducting paired basin studies.

2. There is a need to improve our understanding of the effects of climatic variability and climate change on forest hydrology. Canada covers several physiographic regions that vary in the direction of their long-term climatic trends as well as the amplitude and frequency of oscillations about those trends. Furthermore, the climatic conditions range from those where $P > PET$ (e.g., sites in the Pacific Maritime, Montane Cordillera, Boreal Shield and Atlantic Maritime) to those where $P < PET$ (e.g., sites in the Boreal Cordillera and Boreal Plain). We need to quantify climatic variability and climate change among the forest regions of Canada, and to continue to define the interactions among climate, hydrology and biogeochemistry prior to making effective comparisons of hydrological processes in Canada's various forest regions.

3. There is a need to continue basin-scale monitoring programs in the forest regions of Canada. Canada has several ongoing basin-scale monitoring programs that have been run by provincial and/or federal government agencies for 20+ years (e.g., Turkey Lakes Watershed in Ontario) and that have formed the basis of internationally recognized research (e.g., see special issues on research at the Turkey Lakes Watershed - *Ecosystems* 4, 2001; *Water, Air, and Soil Pollution: Focus* 2, 2001). Unfortunately, the future of these monitoring programs is uncertain as institutional priorities are continuously changing. There are several points to be considered when extending these monitoring programs into the 21st century:

a. Management and science must be connected so that adaptive ecosystem management strategies can be studied. A closer relationship between agencies that are monitoring basins with universities that are conducting process-based studies will have synergistic effects. Long-term databases provide a context for the short-term studies conducted by most university research programs. In turn, process-based studies provide government agencies with greater insight into the hydrological behaviour of the basins, thus assisting in the interpretation of trends in the long-term data.

b. Process-based monitoring and modeling should be embedded in basin monitoring programs. The concept of equifinality suggests that "given the limitations of both our model structures and observed data, there may be many representations of a basin that may be equally valid in terms of their ability to produce acceptable simulations of the available data" (Beven 2000, p. 22). In other words, there is a risk of obtaining the "right" answer for the "wrong" reason. This risk is amplified by focusing solely on external data (e.g., basin streamflow). To place constraints on equifinality, both internal (i.e., hydrological processes) and external data should be collected to calibrate and/or confirm a suite of models of basin hydrological response.

c. We must shift our focus from "average" responses to "spatially and/or temporally distributed" responses. Currently, monitored hydrologic response variables are often aggregated. For example, we often focus on the average yield through time rather than the extremes, and on the average basin yield rather than the spatially- and temporally-variant nature of the contributing source areas of that yield. This limits our ability to assess basin responses to forest disturbance. The focus on average water yield does not allow us to assess, for example, the downstream impacts of increased peak discharges during extreme events, a research issue highlighted in a recent commentary by DeWalle (2003). Secondly, the spatially- and temporally-varying runoff source areas must be identified to develop forest management plans that avoid or minimize disturbance in these areas. This in turn can help mitigate the impacts of forestry operations such as roads, skidder trails and landings on changes in slope stability and on water, nutrient, sediment, and thermal fluxes to receiving waters. Recent technological innovations, including airborne laser altimetry and airborne and satellite remote sensing, should be used to improve our capacity to conduct spatially- and temporally-intensive monitoring.

d. We need to *complete* a network of basin monitoring programs (*sensu* the US Long Term Ecological Research [LTER] sites) that monitor both internal (within the basin) and external (from the basin to the atmospheric or aquatic systems) hydrologic data in significant forest regions in Canada. This will create a valuate resource for both pure and applied hydrological studies in Canada.

e. We need to *coordinate* our network so that common approaches can be adopted to facilitate intra- and inter-basin comparisons.

f. Finally, and probably most importantly, financial and logistical support for these basin monitoring programs must become a priority for government agencies.

4. There is a need to conduct basin experiments within the context of these basin monitoring programs. Basin-scale experiments (or ecosystem experiments) are powerful in that they measure the basin-scale response to forest disturbances, and thus allow managers and scientists to test hypotheses about controls and management of basin processes (Carpenter 1998). Basin-scale experiments are being conducted in Canada; however, recommended improvements in their experimental design include:

a. Inclusion of hydrology (both the science and the scientists) at the planning stage of basin experiments and not after the fact. In this review, the majority of studies reported on basin experiments where different forest disturbance scenarios were investigated by inferring a hydrological role in the observed basin responses without providing data to support these inferences.

b. Explicit consideration of how long the basin should be studied prior to initiation of the "experiment". Experiments based on a "paired basin" design (where one basin serves as the reference and the other as the treatment) are limited, since basins are rarely equivalent in terms of their hydrological features. Experiments based on a "before-and-after comparison" design are limited since the "pre-experiment" data are collected for a time interval that may not fully represent the hydrological conditions that occur during the experiment. Incorporation of process-based understanding (in the form of conceptual or mechanistic models) into the interpretation of pre- and post-experimental data will help alleviate erroneous conclusions based on a "false" reference conditions.

c. Returning to the original spirit of "adaptive management" (*sensu* Holling 1978). Our experimental designs need to be able to respond quickly, both to the insights we obtain into the dominant processes identified in the monitoring/modeling phase and the reality of frequent and/or rapid changes in policies related to forest management (e.g., by the time a basin experiment is completed, the policy that was being investigated may no longer be relevant).

5. Some of the questions that experimental basin research should address include:

a. Does harvesting emulate wildfire? Forest management plans are shifting towards models based on natural wildfire regimes (Hunter 1993). These models assume that fire and harvesting have similar impacts on ecosystems, and that forest management plans that emulate wildfire should preserve ecosystem integrity. These assumptions remain largely unverified (Carignan and Steedman 2000).

b. What is the form and rate of hydrologic recovery following forest disturbance in various forest landscapes in Canada? Studies of this question should recognize that different aspects of a basin's hydrology may recover at different rates.

c. Are the measures that we currently use to protect water resources from the impacts of forest harvesting (e.g., buffer strips along streams and around lake shores) effective in mitigating the hydroecological impacts of harvesting on receiving waters?

d. How do we scale studies at the basin scale to the regional scale? The scale of process studies (< 1 km^2) is significantly smaller than the scale of forest management plans or natural disturbance regimes (> 100 km^2).

Scientists cannot conduct process studies at these coarser scales. We need to develop a scientific basis for scaling processes from the plot or basin scale to the regional scale.

e. What are the cumulative impacts of natural and/or anthropogenic disturbances at the regional scale? The nature of disturbances and their timing, magnitude, and frequency of occurrence at the regional scale may be highly variable. We need a scientific basis for predicting the cumulative impacts of these disturbances on forest hydrology.

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Introduction

The objective of this paper is to highlight the advances made in the hydrology of Canadian wetlands between 1999 and 2003, following a similar report for the preceding four years ((Price and Waddington, 2000). Canadian wetlands are categorized as bogs, fens, swamps, marshes and shallow open water (NWWG, 1997), and comprise 14% of Canada's land area ((National Wetland Working Group, 1988)). Peatlands represent over 90% of Canadian wetlands (Tarnocai, 1998) and that is reflected in the focus of research reported herein. While the focus of this paper is to report on activities between 1999 and (early) 2003, where necessary we have drawn upon other literature on wetlands both within and outside of Canada, to provide the context for more recent initiatives.

Wetlands are areas with the water table at, near or above the land surface for long enough to promote hydric soils, hydrophytic vegetation, and biological activities adapted to wet environments (NWWG, 1997). Wetlands may be mineral-soil wetlands or peatlands, depending on hydrological processes resulting from water exchanges dictated by climate and landscape factors. Mineral-soil wetlands, which include marsh, shallow water, and some swamps, produce little or no peat, because of climatic or edaphic conditions (Zoltai and Vitt, 1995). Peatlands are defined as wetland areas with an accumulation of organic sediments exceeding 40 cm, and include bogs, fens, and some swamps (NWWG, 1997). Fens and (some) swamps are mineratrophic peatlands, receiving water and nutrients from atmospheric and telluric sources, whereas bogs are ombrotrophic, receiving water and nutrients dominantly from direct precipitation. Wetlands exist in the landscape where the water balance ensures an adequate water supply at or near the surface. Thus wetlands are restricted to locations where, on average, precipitation exceeds evaporation loss, or where sustained inflows from surface or subsurface sources alleviate the water deficit. In bogs (by definition) the source is ombrogenous (Woo, 2002), but other wetland types may have a much more complex array of sources.

Considerable progress has been made in wetland hydrology since Ingram's (1983) seminal paper, and is summarized in a recent review by Woo (2002). Within the Canadian context, regional overviews have been provided for Quebec and Labrador (Price, 2001); the McKenzie Basin (Rouse, 2000); the Arctic (Woo and Young, 2002); and the Prairies (van der Kamp and Hayashi, 1998; LaBaugh *et al*., 1998).

Atmospheric Fluxes and Stores

Quantification of the component fluxes and stores continues to be an important first step in understanding, managing and modeling of wetlands. Precipitation provides the source of water directly to wetlands and for recharging surface and groundwater inputs, and varies tremendously across Canada. Precipitation, particularly snow inputs and interception losses, can be difficult to quantify. This is especially true in large watersheds because the month-to-month changes in snow-cover, for example, are often much greater than the atmospheric flux terms (Strong *et al*., 2002). The input must overcome interception losses, which in forested wetlands may represent 32% for black spruce (Van Seters and Price, 2001); or 21-25% for larger cedar and balsam fir forests in BC, with only 1% being stemflow. Fog interception was shown to be notable, if not significant, in a hypermaritime northwest BC wetland forest (Emili and Price, unpublished data).

More attention has been given to evapotranspiration fluxes from wetlands. Land surface characterizations such as CLASS (Canadian Land Surface Scheme) (Verseghy, 1991), for use in global climate models, recognize the importance of wetlands. Bartlett *et al.* (2002) noted that overrepresentation of organic soils in heterogeneous environment results in the overestimation of the evaporation loss. Surfaces dominated by vascular plants are well represented, but non-vascular surfaces (e.g. *Sphagnum*) remain problematic (Comer *et al*., 2000). Given that latent heat losses are generally the largest water sink in wetlands, research continues into the role of soil and vegetation on the energy balance. Eaton and Rouse (2001), for example, found that over a 10-year period the seasonal variability of cumulative evapotranspiration from a subarctic sedge fen was controlled primarily by cumulative precipitation. Petrone *et al.* (in press, a) found that the presence of a higher water table, higher soil moisture and the presence of vascular plants in a cutover peatland being restored, resulted in a higher evapotranspiration loss than an adjacent unrestored site. The emergence of vascular vegetation (birch) increased the surface roughness over a single growing season (Petrone *et al*., in press, b).

Surface & Groundwater Fluxes

By virtue of the location of wetlands within the landscape, understanding wetland hydrology requires knowledge of groundwater and surface water interactions. The proportion of surface and groundwater inputs, and a wetland's interaction with groundwater (i.e. recharge vs. discharge function) is governed by its position within the groundwater flow system, the hydrogeologic characteristics of soil and rock material, and their climatic setting (Tóth, 1999; Winter, 1999; Sophocleous, 2002). The extent and type of groundwater interaction will influence the amplitude and duration of baseflow and water level fluctuations, which is linked to surface saturation, runoff processes and ultimately sediment redox, biogeochemical process, and vegetation patterns within a wetland (Tóth, 1999; Hill, 2000; Hayashi and Rosenberry, 2001).

There has been considerable research directed towards determining the role of groundwater- surface water interactions in wetland hydrology from a variety of climatic and geologic regions throughout Canada. In permafrost dominated areas of northern Canada, groundwater – surface water interactions are restricted to localized flows in the surface active layer during periods of thaw (Woo and Young, 2003). In such arid climates, patchy arctic and subarctic wetlands occupy low areas inundated during the snowmelt period, or with water storage sustained by more gradual inputs from groundwater or late-lying snowbanks (Carey and Woo, 2001a; Woo and Young, 2003). The shallow frost table limits percolation thus facilitates the existence of these systems (Carey and Woo, 1999; 2001b; Young and Woo, 2002). Coarse sorted stones between frost mounds (Hodgson and Young, 2001), or macropores and pipes between peat hummocks (Carey and Woo, 2000; 2002; Quinton and Marsh, 1999), supply water to wetlands. Wetlands such as these, having a strong dependency on water from localized and shallow sources like supra-permafrost water, ground-ice melt and late-lying snowbanks, are susceptible to climate change (Young and Woo, 2000), particularly in view of the expected deepening active layer and large evaporative losses (Young and Woo, 2003).

In topographically controlled humid eastern boreal regions with shallow soils underlain by impermeable crystalline bedrock, groundwater is usually restricted to localised flow, resulting in predictable and relatively simple wetland groundwater interactions. Shallow near surface runoff dominates and flow rates are inextricably tied to shorter-term variation in weather and, thus, susceptible to extended dry periods and climate change (Devito *et al.,* 1999a). Although groundwater is a minor input to headwater shield wetlands, soil depth within the catchment can influence the seasonality of groundwater connectivity between uplands and wetlands influencing water table fluctuations and maintaining surface saturation during drought, increased runoff during summer storms (Devito *et al.,* 1999a; Hill 2000).

In regions of deeper glacial deposits, such as the Great Lakes and Laurentian regions of humid eastern Canada, larger scale groundwater interactions occur as influenced by topography and substrate grain size distribution. Groundwater interactions with wetlands located in finer grained glacial tills are limited by low permeability. Similar to Precambrian systems, catchment storage is low and near surface runoff during storm events dominate the inputs and outputs (Hill, 2000; Vidon and Hill, 2003). These wetlands are maintained in depressions and breaks in slope with ample supply of water in humid climate, but susceptible to drought periods (Hill, 2000). However, the role of larger scale topographically controlled intermediate or regional groundwater flow, or heterogeneous lithology (sand lenses) in clay-rich glacial deposits on wetlands in this region is poorly understood. In adjacent outwash landscapes with permeable surface aquifers of coarser grained material, constancy of groundwater inputs, connection, permanence of surface saturation and stability of water table in wetlands was observed with increasing depth of aquifer (Hill, 2000; Vidon and Hill, 2003; Warren *et al.,* 2001). Aquifer lithology and deposits of lower hydraulic conductivity peat in near-stream wetland areas can also influence wetland groundwater–wetland interactions, surface saturation and runoff in glaciated outwash regions of Ontario (Devito *et al.,* 2000a; Hill *et al.,* 2000). Reversals in hydraulic gradient and groundwater flow from streams into adjacent wetlands, or wetlands back into the hillslope appears to be influenced by riparian wetland slope and aquifer depth in outwash landscapes (Vidon and Hill, 2003). Groundwater reversals across a barrier beach also occur when hydraulic gradients controlled by lake and/or marsh water level change seasonally (Huddart *et al.*, 1999).

In the dry Prairies and Boreal Plain of western Canada regions, external water inputs become increasingly important to wetland maintenance. Groundwater flow is complex due to the low relief and deep glacial deposits and depending on their position, wetlands can have recharge, flow-through or discharge function (van der Kamp and Hayashi, 1998; LaBaugh *et al.,* 1998; Tóth, 1999; Devito *et al.,* 2000b). However, in clay rich glacial deposits groundwater interaction is a minor component of most Prairie wetlands due to the low permeability (Conly and van der Kamp, 2001; van der Kamp *et al.,* 2003; Parsons *et al.* , in press). Although soil storage is low in clay –rich tills, storage is rarely exceeded during the s}mmer, and surface runoff into ponds during this time is limited to infrequent large storm events. Thus, ponds and depressions are heavily reliant on snowdrift and snowmelt runoff over frozen soils and experience considerable seasonal water-level variability (van der Kamp *et al,.* 1999; Hayashi *et al.*, 2003). These systems are generally considered susceptible to changes in climate and landuse that influences snowmelt surface water interactions (van der Kamp *et al.,* 1999; 2003). Depressional wetlands such as these are focal points for groundwater recharge (Hayashi *et al.*, 1998; 2003), though most of the water recharge may flow to the moist margins, rather than to deep percolation (Parsons *et al., in press*). Such lateral recharge may be important in sustaining water yield to local shallow wells (Hayashi and van der Kamp, 1998). The storage function of depressional wetlands can be represented mathematically with volume-depth-area relationships (Hayashi and van der Kamp, 2000; Weins, 2001), which can be incorporated into a storage-water balance model (Su *et al.,* 2000) that can simulate long-term water level variations (Conly and van der Kamp, 2001). However, comparison of pond systems located in coarse-grained glacial deposits in the Canadian Prairies is lacking, and regional variation in groundwater interactions on depression wetlands in coarse-grained glacial deposits is poorly understood (Winter, 1999; 2001).

Similar to Prairies systems, groundwater exchange to Boreal Plain pond-peatland complexes located in topographic high and low areas of fine textured, low conductivity glacial till and lacustrine deposits contribute little to the water balance (Ferone and Devito, in press). Further, forested clay-rich hillslopes provided little snowmelt or storm runoff in most years because soil storage and transpiration demands exceed rainfall, that occurs primarily during the summer in this sub-humid environment (Devito *et al.*, submitted; Kalef, 2002). In contrast to semi-arid Prairies, runoff generation is primarily from near surface flow from peatland areas connected to shallow pond wetlands that develop in sub-humid climate (Wolniewicz,2002). Topographically high ponds act as focal points for recharge, primarily providing lateral flow from the pond to peatland and adjacent hillslopes (Ferone and Devito, in press). During larger rain events, the hydraulic gradient between peatland and pond reverses and peatland discharges water into the pond. The low lying pond-peatland complex functioned as a flow through system of near-surface water originating from extensive peatlands adjacent to the pond (Ferone and Devito, in press). The differences in shallow groundwater interaction and topography also influenced pond chemistry (Ferone, 2001) and have implication for pond response to climate or disturbance (Tóth, 1999; Winter, 2001). Further, comparative studies of ponds located in adjacent landforms dominated by coarse-grained glacial deposits (outwash sands and gravels), indicates a significant exchange of groundwater from larger scales of flow on pond water budgets, and different responses in water levels and impacts to climate and disturbance (Halsey and Devito, submitted; K. Devito, unpublished data).

The boreal and temperate mountain and plains areas within in the Western Cordillera provide excellent locations for conducting field studies to increase our understanding of processes influencing groundwater-surface water interactions in wetlands from a range of topographic, geologic and climatic conditions. Considerable work has been conducted on groundwater interactions in forested and montane uplands and streams (Carey and Woo, 1999, also see Buttle *et al.*, this issue). However, initial studies on groundwater – surface water interactions within wetlands have been conducted in this region since the last review. Wetlands and lakes have been shown to moderate the thermal regimes of groundwater fed streams draining cut and forested catchments in the interior of British Columbia (Mellina *et al.,* 2002). Fitzgerald *et al.,* (in press) showed small headwater swamps in north coastal BC are critical interfaces between steep, well-drained forested upland slopes and runoff, and should be avoided during timber harvesting. Surface runoff from sloping forested swamps on moderate slopes organizes itself quickly into “seeps” (Emili and Price, unpublished data), which diverts this channellized water away from more gently sloping blanket bogs (Fitzgerald *et al.,* 2003). Runoff from these blanket bogs thus has a much lower runoff ratio than micro-catchments containing these seeps (Emili and Price, unpublished data).

Internal water flow through peat wetland systems has received relatively limited attention recently. Reeve *et al.* (2000) suggested the extent of vertical flow in peatlands such as in the Hudson Bay Lowland, is limited by low permeability mineral substrate (i.e. rather than the humified peat layers), promoting lateral flow through the acrotelm. However, Fraser *et al.* (2001a) noted reversals between recharge and discharge to and from the deeper peat layers during wetter and dryer periods, respectively. Similarly, vertical flow has been observed in Boreal Plain peatlands during extended periods dry periods (Halsey and Devito, submitted). Tranfer of water from deeper peat layers to the surface is important in sustaining soil moisture in the surface layer, and changes of soil moisture in this layer should be accounted for in the water balance (Lapen *et al*., 2000). Vertical redistribution of peat pore-water has also been shown to occur from the saturated zone to the soil-moisture zone (Kennedy, 2002) during peat consolidation (Lang, 2002), and may be enhanced by pressure caused by methane generation (Price, 2003). These exchanges have implications for water quality (Fraser *et al.,* 2001b).

Runoff

Runoff from wetlands is controlled by the rate and magnitude of inputs, and the efficacy of storage, usually involving an interaction of surface and shallow groundwater (Gibson *et al.,* 2000). While the presence of surface water suggests its importance as a delivery mechanism, mixing model studies demonstrate the predominance of “old” water in storm runoff from headwater swamps, for example (Brassard *et al*., 2000; Fitzgerald *et al.,* 2003). The efficiency of these mechanisms was reflected by higher runoff to northern Alberta lakes in wetland-dominated catchments, compared to upland systems (Gibson *et al*., 2002). The expansion and connectivity of saturated surfaces associated with peatland areas in the boreal plains region of Alberta are directly related to regional runoff regimes (Wolniewicz, 2002). Where deep seasonal frost or permafrost is present, frost-table dynamics control the runoff pathways (Quinton and Marsh, 1999; Carey and Woo, 2001b) and “new” water (snowmelt) may dominate (Metcalfe and Buttle, 2001). In contrast, wetlands with deeper flow systems may experience a notable baseflow component (Beckers and Frind, 2001). In spite of the efficiency of wetland runoff processes, the flood mitigation role of wetlands is often overstated (Simonovic and Juliano, 2001). Modelling of wetland runoff-response (McKillop *et al.,* 1999) requires careful parameterization of hydraulic parameters at a variety of scales (e.g. Letts *et al.,* 2000).

Water Flows in Disturbed Wetlands

The health of wetland systems is threatened by various direct and indirect anthropogenic changes (Detenboeck *et al.,* 1999). Anticipated changes in the global climate are expected to increase the soil-water deficit in North America, which will alter the water balance of wetlands. Monitoring networks often have insufficient spatial and temporal sampling to establish patterns of variability or provide insight into processes related to climate change (Conly and van der Kamp, 2001). At the local scale wetland disturbances may be direct and intentional, such as for agriculture (van der Kamp and Hayashi, 1998), drainage of forestry (Prevost *et al.*, 1999), or peat mining (Price *et al.,* 2003).

Forest drainage following harvesting of a treed peatland has been shown to improve growth of naturaly regenerated black spruce (Jutras *et al.,* 2002). While short-term studies concerning planted seedlings cannot effectively corroborate this (Roy *et al.,* 2000b), there are some limitations to the effective lowering of the water table. This arises because of changes to the peat substrate following drainage, including subsidence and changes in aeration dynamics (Silins and Rothwell, 1999), accelerated peat decomposition (Prevost and Plamondon, 1999), and a decrease in the effective hydraulic conductivity as the water table drops (Belair et al. 2003). There is also concern that "wetting up" caused by decreased interception (Roy *et al.,* 2000a), may impair root aeration and accelerate *Sphagnum* growth sufficiently to overtake seedlings (Roy *et al.,* 2000b). Drainage at these sites increased basefow, caused a 25% increase in total runoff, increased dissolved and suspended solids, and temperature variability in outflow (Prevost and Plamondon, 1999).

Sites that are mined for peat have additional stresses. Removal of the living layer, or acrotelm, severely impairs their hydrological function (Price *et al.,* 2003). Following abandonment, drainage ditches can remain effective for decades (Van Seters and Price, 2000), and the irreversible changes to morphology and peat structure alter flow patterns (Van Seters and Price, 2001). Spontaneous regeneration of the dominant peat-forming moss, *Sphagnum*, is curtailed because of the unstable surface caused by needle-ice formation (Groeneveld, 2002), the strong capillary retention of the cutover substrate and development of a litter layer of ericaceous leaves (Price and Whitehead, 2002), which restrict water flow to the mosses. Regeneration in wetter areas of old manually block-cut peatlands can occur spontaneously (Whitehead and Price, 2000). While average wetness conditions on abandoned vacuum-harvested sites (the modern extraction technique common in Canada) are little different than block cut sites, the extreme spatial variability of wetness in the latter sites ensures that at least some loci for regeneration exist (Price *et al.,* 2003). Restoration plans must account for the cumulative human impacts on wetlands, including broader landscape effects (Bedford, 1999). Managed restoration requires rewetting of the site, often by blocking ditches. More aggressive restoration management has demonstrated the effectiveness of artificial terraces (von Waldow, 2002) or shallow basins (Price *et al.*, 2002) to retain snowmelt water. Application of straw mulch reduces evaporative losses (Petrone *et al.,* 2001; in press a), and reduces the scale of soil moisture variability (Petrone *et al.,* in press b). Important changes to the hydraulic character of the remaining peat occur with drainage and peat extraction (Schlotzhauer and Price, 1999), that profoundly affects storage exchanges (Price and Schlotzhauer, 1999). Compression of the peat caused by seasonal declines in the water table (Lang, 2002) can cause up to a three order of magnitude decline in hydraulic conductivity (Price, 2003). Incorporation of these peat volume changes and its affect on hydraulic parameters, into a vertical 1-D numerical flow model (Kennedy, 2002), allowed the prediction of soil-water pressure, soil moisture, water table and peat surface elevation, and prediction of various abandonment and restoration scenarios.

At the local to regional scale, indirect disturbance can impact or interact with hydrological cycle and surface and groundwater linkages between adjacent uplands and wetlands. These linkages are important for maintaining the hydrological and ecological integrity of wetlands (*eg.* Hayashi and Rosenberry, 2001). Such interactions fall within the realm of forest or upland hydrology and the reader is referred to Buttle *et al.* (this issue). Clearly a broader catchment /landscape approach is required to understand upland-wetland linkages and the potential impact of individual and cumulative catchment disturbance to the receiving wetland (Hill, 2000; Bedford, 1999). Examination of wetland position within the surface and groundwater flow system can indicate the relative importance of regional and local scale disturbance impacts on wetlands and other aquatic systems (Hill, 2000; Devito *et al.,* 2000b). Limited runoff from both forested and cut portions of a sub-humid boreal catchment, due to the low rainfall relative to soil storage and evapotranspiration, resulted in little influence of logging aspen uplands on the hydrology of groundwater fed valley wetland (Kalef, 2002). The conversion of cultivated fields to permanent grass and shrub resulted in the drying-out of adjacent depression wetlands (sloughs) by severely reducing localized snow drift and spring melt runoff inputs (van der Kamp *et al.,* 1999; 2003).

Carbon Cycling

Undisturbed peatlands are presently a relatively small sink for CO_2 and a large source of CH_4. When the 'global warming potential' of CH_4 is factored in, many peatlands are neither sources nor sinks of greenhouse gases (Waddington and Roulet, 2000). However, land-use change significantly alters greenhouse gas emissions (Roulet, 2000). Considerable progress has been made on understanding carbon cycling processes in Canadian mined, cutover and restored peatlands.

The exploitation of peatlands for *Sphagnum* is widespread in North America and Europe. Peat mining, through the combination of drainage, peat removal and subsequent abandonment, alters the environment so severely that *Sphagnum* spp. are unable to colonize. Cutover peatlands represent a persistent source of atmospheric CO_2, losing 300 to 400 g C m^{-2} yr^{-1} (Waddington et al., 2002; Waddington and McNeil, 2002; Petrone *et al.,* in press). Peat oxidation is more dependent on peat moisture content and peat temperature (Waddington *et al.,* 2002) than peat carbon quality (Waddington *et al.,* 2001) suggesting water management (e.g., Price *et al.,* 2002) may reduce CO_2 losses from mined peatlands (Waddington and Price, 2000).

The restoration of *Sphagnum* mosses on cutover sites has the potential to sequester atmospheric CO_2 thereby returning the peatland to a peat accumulating system (Waddington *et al.,* 2003). Restoration not only increases plant production but also decreases total respiration (Waddington and Warner, 2001). *Sphagnum* production varies between species (e.g., S. *fuscum* > S. *capillifolium*) according to their ability to withstand harsh conditions on restored peat surfaces (Waddington *et al.,* 2003a). A stable moisture supply is more beneficial to *Sphagnum* growth (Rochefort *et al.,* 2002) compared with repeated wetting and drying events (McNeil and Waddington, 2003). The application of a mulch surface improves moisture conditions near the peat surface but mulch decomposition represents a short-term source of atmospheric CO_2 (Waddington *et al.,* in press). Petrone *et al.* (2001) determined that a recently restored peatland was a larger source of CO_2 than an adjacent cutover site in part because the mulch decomposition exceeded the new production of mosses and vascular plants.

As vascular plants colonize restored peatlands CH_4 flux increases due to both the supply of labile carbon and the enhanced CH_4 transport (Day and Waddington, unpublished data). CH_4 flux from cutover peatlands is greatest in drainage ditches (Waddington and Price, 2000) owing to their permanently flooded conditions and supply of highly labile DOC (Tóth, 2003). Belissario *et al.* (1999) found sites of high CH_4 emissions in undisturbed peatlands had enriched $\delta^{13}CH_4$ signatures suggesting the importance of acetate fermentation pathway on methanogenesis. DOC has also been correlated to CO_2 exchange in cutover peatlands (Glatzel *et al.,* in press).

Many landscape scale studies of carbon exchange from undisturbed peatlands have taken place in Canada in the last few years. These and many carbon cycling studies have been constrained to growing season measurements. However, Lafleur *et al.* (2001a) determined that the non-growing season CO_2 loss from an ombrotrophic peatland is not small (183 g CO_2 m^{-2}). On an annual basis, however the peatland was an annual net CO_2 sink (248 g CO_2 m^{-2} yr^{-1}). Joiner *et al.* (1999) determined that a boreal fen was a net source of CO_2, losing 31 g C m^{-2} in 1994 but was a net sink in 1996 (-92 g C m^{-2}). The inter-annual difference was linked to an earlier snowmelt and thaw of the fen surface – leading to drier summer conditions. Griffis *et al.* (2000a) suggest that an early snowmelt combined with wet and warm conditions during the spring period leads to large carbon acquisition even when drier conditions prevailed over the majority of the growing season. CO_2 exchange in an adjacent wetland forest, however, was related to timing of snowmelt and heat content prior to leaf out (Lafleur *et al.* 2001b; Rouse *et al.,* 2002).

Peatland surface topography leads to differences in carbon exchange processes and vegetation assemblage (Waddington and Roulet, 2000). Griffis *et al.* (2000b) scaled community-level CO_2 measurements from hummocks and hollows, to tower measurements at the landscape scale. Variability in peatland surface topography (hummock, hollow, peat platueau, etc.) leads to differences in soil moisture, temperature, vegetation type and biomass (Moore *et al.* 2002). This creates differences in peat CO_2 production (Scanlon and Moore, 2000), the fluxes of CH_4 and CO_2 (Dalva *et al.,* 2001), and peat accumulation (Robinson and Moore, 1999). It is expected that different communities will respond differently to climate change (Griffis *et al.,* 2000b). Clair *et al.* (2002) suggest that under a 2x CO_2 climate change scenario that carbon loss from a small temperate wetland will almost double from 0.6% to 1.1% of total biomass.

Differences in vascular and nonvascular vegetation-carbon-water dynamics were incorporated in a dynamic model of long-term peat accumulation (Frolking *et al.,* 2001). The model suggests that bogs are more sensitive than fens to climate conditions. Moreover, warmer and wetter conditions were found to be more conducive to peatland development (Frolking *et al.,* 2001). In a simpler model focusing on the non-linear interactions among peat production, decomposition and hydrology, Hilbert *et al.* (2000) also demonstrate the sensitivity of peat accumulation to peatland water balance.

The Peatland Carbon Simulator (PCARS), developed by Frolking *et al.* (2002), is a process-oriented model of the contemporary carbon balance of northern peatlands. Seasonal patterns and the general magnitude of net ecosystem exchange of CO_2 were similar with measured tower data (see Lafleur *et al.* 2001a). PCARS was designed to link to the CLASS land surface model and will prove valuable in examining climate-peatland carbon feedbacks in future research. The model incorporates the exchange and interaction of CH_4, CO_2, and DOC.

Fraser *et al.* (2001b) determined that DOC export from an ombrotrophic bog was 12% of the magnitude of the carbon sink measured at the same peatland (Lafleur *et al.,* 2001a). DOC concentration in the acrotelm was variable and 'allochthonous-like', whereas catotelmic waters were more 'autochthonous-like'. The influence of hydrology on the patterns of supply and quality of DOC has also been shown to have a major influence on the cycling of nitrogen (Hill *et al.,* 2000) and mercury.

Water Quality

Process-level investigation of the hydrological and biogeochemical controls on the mechanisms and rates of element transformation (Hill *et al*., 2000; Devito *et al*., 2000a) are some of the most important advancements in the understanding of water quality in wetlands. For example, in a forested riparian wetland that received groundwater with elevated nitrogen concentration (10-30 mg N L^{-1}), denitrification was carbon limited and only became nitrogen limited in narrow zones of strong denitrification in pockets of buried peat (Hill *et al.,* 2000). These findings were supported through nitrogen isotope analyses (Devito *et al*., 2000a), confirming that groundwater flow paths, and the adequate supply of terminal electron donors and acceptors, control microbial denitrification in riparian wetlands. During snowmelt periods when surface flow occurs, runoff can bypass riparian "buffer" zones and directly enter streams (von Waldow *et al*., 2000). Isotopic methods combined with *in situ* tracer experiments can demonstrate microbial denitrification activity and rates (Mengis *et al.,* 1999). The spatial discontinuities in denitrification zones controlled by hydrology and lithology can confound denitrification mechanisms (Hill, 2000). Catchment geomorphology and hydrological connectivity must be considered when assessing catchment nitrogen dynamics (Devito *et al*., 1999b; Shiff *et al*., 2002).

Phosphorus dynamics remain poorly understood (Devito *et al*., 2000b). Carlyle and Hill (2001) determined that groundwater flow governed redox conditions, and geochemistry strongly influences the solubility and mobility of phosphorus. A detailed three dimensional analyses of groundwater flow, dissolved oxygen and iron species (Carlyle and Hill, 2001) revealed anaerobic zones where Fe^{3+} was reduced to Fe^{2+}. This influenced groundwater soluble reactive phosphorus (SRP) concentrations. As with denitrification (Hill *et al.,* 2000), patterns of SRP concentration reflect local interaction of flowpaths with different redox states.

Microbially-catalyzed redox reactions have recently been investigated in more detail (Branfireun *et al*., 1999). Laboratory incubations by Blodau *et al.* (2002) showed iron cycling in northern peatlands had little influence on carbon flow, but sulphate reduction had the potential to limit methane production by 48 to 86%. Fortin *et al.* (2000) found significant activity of both iron and sulphate reducing bacteria in a young constructed wetland during the winter while Kennedy and Mayer (2002) noted that cold weather performance must be better understood before water- treatment wetlands can gain widespread acceptance in Canada.

Wetlands remain a focus of research in mercury cycling, both in terms of the mechanisms governing the production of methylmercury *in situ* (Branfireun *et al.,* 1999; Heyes *et al*., 2000) as well as their controls on the fate and transport of both inorganic mercury and methylmercury (Branfireun and Roulet, 2002; Rencz *et al*., 2003; Young and Branfireun, submitted). At the mesocosm scale, Branfireun *et al.* (1999) investigated the link between atmospheric sulphate deposition and the production of methylmercury in peat. They found a clear increase in the amount of methylmercury, a potent neurotoxin, with sulphate addition, making the first direct link between sulphate deposition in precipitation and the mercury cycle in peatlands. Most research on mercury cycling in wetlands has focused on acidic, nutrient-limited S*phagnum* dominated peatlands. However, Young and Branfireun (submitted) found that the transport of total mercury in a temperate swamp was greatest when the wetland and surface streams were hydrologically connected. Methylmercury was released downsteam, with yields similar to that of more acidic peatland systems. An inverse relationship between methylmercury and sulphate concentrations was observed, with the highest and lowest concentrations, respectively, found during the periods of persistent inundation and the onset of anaerobic conditions in the wetland sediments.

The observation of an inverse relationship between inundation and sulphate concentrations are consistent with the findings of Warren *et al.* (2001). They found a strong relationship between hydrological connection, reducing conditions in temperate swamp sediments and stream sulphate concentrations (Eimers and Dillon, 2002; Devito *et al.,* 1999a). This clarified the inverse relationship between water level fluctuations and sulphate reduction and oxidation (Devito and Hill, 1999).

Studies of the effect of anthropogenic impacts on wetland water quality are relatively sparse. Heyes *et al.* (2000) found that elevated methylmercury concentrations was the consequence of Hg methylation in an experimentally flooded anaerobic wetland surface. The flooding facilitated the exchange of nutrients, importantly sulphate, between the peat surface and the surface water. Prévost and Plamondon (1999) found surface water from drained sites had significantly higher electrical conductivity than natural peatlands because of leaching of N, Na, S, Ca, and Mg. Water quality did not return to pre-treatment levels even after five years. The effects of atmospheric contaminant loading to wetlands via precipitation may significantly influence biogeochemical cycles (Branfireun *et al.,* 1999) or ecosystem function (Donald *et al.,* 1999; 2001). Donald et al. (1999) found that the concentrations of agricultural pesticides in wetlands in the Prairies were positively related to precipitation amount, and often exceeded the limits recommended for the protection of aquatic biota.

Scaling Issues

Generalizing site specific studies and scaling up from the upland-wetland boundary to the regional scale is a major research challenge, not only for wetland hydrology (Sophocleous, 2002). Devito *et al.* (2000) developed a hierarchy of landscape factors predicting surface and ground water connectivity of wetlands and lakes, and the potential susceptibility of surface waters to land-cover changes. Such qualitative landscape approaches may need to precede more quantitative modelling methods (e.g. Woo and Young, 2003). Currently, regional 3-D surface and groundwater hydrological models are not well equipped to represent more local phenomena at the interface of groundwater – surface water interactions in wetlands (Sophocleous, 2002). Use of readily available surficial geology data, topography and climate as indices of hydraulic gradient and permeability provide a reasonable basis for prediction of the dominant components of the water balance and the scale of linkages of the wetland to the surrounding a region.

Advances in remote sensing of wetlands has provided a method for generalizing site specific to regional studies and assessing dominant hydrological processes at different scales. For example, RADARSAT, LANDSAT and SPOT images provide data for assessing historical and current patterns of flooding in the Peace-Athabasca Delta (Toeyrae *et al.,* 2001; 2002), and for parameterization of hydraulic flow models (Pietroniro *et al.,* 1999). Such methods are suited to evaluation of aggradation or plant succession in these systems, and may clarify or challenge the paradigm of delta degradation, such as that attributed to the effect of the WAC Bennett Dam on the Peace-Athabasca Delta, for example (Timoney, 2002). LANDSAT and RADARSAT images can also be used to predict the dominant runoff generating areas, and dynamic nature of small localized depressions and coarse scale catchment boundaries in influencing regional runoff (Wolniewicz, 2002).

Conclusions

Considerable challenges face wetland hydrology and water quality research. Data are often spatially and temporally too sparse to adequately characterize the dynamics of wetland hydrology and especially the biogeochemical processes. Biogeochemical and hydrological studies are often not satisfactorily convergent, and there are profound uncertainties surrounding the impact of environmental change on wetland biogeochemical function (Hill, 2000). Many of these problems are intractable in either the field or the laboratory, but a promising convergence of scales of investigation is emerging that will couple these two approaches, leading to significant advances in the coming years.

Snow, Frozen Soils and Permafrost Hydrology in Canada, 1999 - 2002

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Abstract:
This paper provides an overview of Canadian research on snow, frozen soils and permafrost hydrology for the period 1999-2002; the period between the 1999 IUGG meeting in Birmingham and the 2003 IUGG in Sapporo. During this period there were significant advances in both our understanding of the physical processes, and our ability to model these processes. This report assesses these advances.

INTRODUCTION
Woo et al. (2000) outlined the advances in Canadian snow, frozen soil and permafrost hydrology for the previous 4 year period. During this period, there were advances in understanding of related processes, and the development of models of snow accumulation and melt, relocation of snow by wind, snow interception in forest canopies, and sublimation and energy balance during snowmelt. The influence of heterogeneous topography, vegetation and snow properties were discussed in a number of papers. The current paper will consider the advances made in Canadian snow and frost hydrology, since Woo et al., (2000). We will not be considering snow issues directly related to forest hydrology, remote sensing of snow, or snow chemistry issues. Portions of these issues will be considered in other Canadian IUGG review papers.

SNOW HYDROLOGY
Recent progress in snow hydrology has continued to consider the spatial heterogeneity in surface energy fluxes, and therefore, melt, and the temporal variability in snowmelt. In addition, there has been a renewed emphasis in modeling of snow hydrology and snow ecology. The following will concentrate on these aspects.

Snow Processes and process modelling
Pomeroy et al. (2002) provided simple, yet physically appropriate, equations for estimating snow accumulation beneath forest canopies. This approach relates canopy properties (leaf area index or canopy density) to either snowfall or snow accumulation in clearings. For sites where mid-winter melts, wind redistribution and surface evaporation are infrequent or small, this approach can estimate snow accumulation. Comparisons with data and approaches from eastern Europe and Siberia, suggest that these results are transferable.

Woo and Giesbrecht (2000) presented a model of the effects of a single spruce tree on both the turbulent fluxes and radiation to the snow cover immediately surrounding a single tree. This study illustrated decreased snow water equivalent (SWE) close to the trunk, as well as low turbulent fluxes related to a low wind regime below the canopy, and the dominance of radiation in controlling melt. The tree resulted in reduced shortwave radiation, and increased long wave radiation relative to an open site. These factors, when combined with tree and solar geometry result in large differences in melt with distance away from the trunk, and significant contrasts in melt rates in different directions. Giesbrecht and Woo (2000) utilized the single tree model, to upscale these processes to a stand scale (80 x 64 m^2), with all trees within this area included. However, since this was an open canopy, sub-arctic forest, the tree canopies only covered 2.1% of the study plot area. This model was able to predict the spatial pattern of melt, and the decrease in plot SWE. An important result from this study, is the demonstration of the large spatial variability in melt. For example, they demonstrated that melt from a sunny location in a forest opening can overestimate average melt by over 200%. However, this study did not consider the role of advection once bare patches developed, or the effect of shrub undergrowth on melt rates.

Using a different approach to that of Woo and Giesbrecht (2000), Faria et al. (2000) also considered the relative importance of variable SWE and melt energy for forests (including spruce, pine, mixed, burned, and open) in controlling snow cover depletion in central Saskatchewan. Using detailed snow surveys and measured snow cover depletion throughout the melt period, they also found variable SWE and variable melt rates. They found that melt rate is inversely correlated with SWE for spruce forests (i.e. SWE increased and melt rates decreased with distance from the trees). As a result, snow-free patches first appeared close to the spruce trees, and then expanded in area over time. This covariance between SWE and melt rates was the major factor controlling the depletion of the snow cover. For these sites, where the SWE increases with distance from trunks, it is likely that a combination of radiation from trunks and advection from trunks and bare ground resulted in the covariance of melt and SWE. The result of this covariance is an acceleration in the reduction in snow cover area, compared to that caused by variations in SWE alone. This is similar to the results of Woo and Giesbrecht (2000). However, Faria et al. (2000) also found a covariance between melt and SWE for deciduous, burned and open sites. Possible mechanisms in these cases include decreased albedo in areas of shallow snow due to either greater leaf litter concentration and/or the influence of the forest floor. Another possibility was increased advection of sensible heat from shrubs as the snow became shallower. Iacozza and Barber (2001) showed that for sea ice in the Canadian Arctic, snow melt rate was spatially non-uniform, with melt rates decreasing non-linearly with increasing snow depth until a depth of approximately 16 cm, above which melt rate was fairly uniform. They explained this pattern as a result of lower albedo for the shallow snow sites. In addition, they speculated that larger equi-temperature snow grains occur in the shallower snow due to differences in thermal diffusivity. It was suggested that these larger snow grains also had a lower surface albedo, again resulting in higher melt rates with shallower snow.

Previous work on the advection of heat between snow patches and snow-free patches had utilized detailed boundary layer models (Liston, 1995; Marsh et al., 1999); simplified approaches based on Weismann (1977) (Shook, 1995); or have utilized simple parameterization based on the proportion of the area which is snow covered (Neumann and Marsh, 1998). One deficiency of these methods is the inability of the boundary layer models to properly account of the small scale variability in a natural snow cover or the inability of the simple parameterizations to consider the effects of patch size and geometry on advection (Marsh et al., 1999). Granger et al. (2002) attempted to overcome these issues by developing a boundary-layer integration approach. They developed a simple series of equations to estimate the advected energy term, that reduces the advection term to a simple power function of the snow patch size. This simple model requires standard meteorological parameters, plus surface temperature of the snow and snow-free patches and the snow patch length. Further work is required to test this approach to field estimates of advection.

As noted by Marsh (1999), there have been few observations of the spatial variability in the fluxes of sensible and latent heat over snow covered surfaces. During the Mackenzie GEWEX Study (MAGS) observations of the 1998/99 water year for the Mackenzie River basin, the Canadian National Research Council Twin Otter, carried out observations in the Inuvik region. Brown-Mitic (2001) showed spatial flux maps for a tundra site north of Inuvik. These maps illustrated the large spatial variability in sensible and latent heat fluxes over this tundra site during a period when the basin was partially snow covered. Ongoing work is considering the processes controlling this variability. Brown-Mitic (2001) also showed the large differences between snow covered tundra and boreal forest sites, with the tundra site using more than 80% of the net radiation to non-turbulent fluxes (primarily snow melt and warming of the soil), whereas at the forest site, less than 50% is used for non-turbulent fluxes.

Large Scale studies

The Mackenzie GEWEX Study (MAGS) has considered the influence of snow on the water and energy fluxes of the northward flowing Mackenzie River (Stewart et al., 1998; Stewart et al., 2002; Rouse et al., 2003). As part of a water year study of the Mackenzie River Basin, Stewart et al. (2002) showed the variability of snow cover fraction, days with snowfall, and days with blowing snow, for the entire Mackenzie Basin, and considered both the external and internal factors controlling changes in these conditions for the 1994/95 water year. In the same study, Marsh et al. (2002) showed that the spring of 1995, was the earliest melt on record for the lower Mackenzie Basin, and that there was a trend towards earlier melt in this portion of the Mackenzie Basin.

A version of the Canadian Regional Climate Model (CRCM), has recently been coupled with the Canadian Land Surface Scheme (CLASS), allowing high resolution model for considering mesoscale circulations during the snow melt period over the Mackenzie River Basin (Mackay et al., 2001). This study showed that mesoscale circulations were formed due to heterogeneous surface forcing, including strong sensible heat fluxes over snow-free canopies. These circulations produced mesoscale downdraft regions, that along with adiabatic heating, resulted in higher turbulent fluxes to the snow cover.

Snow modelling

The Canadian hydrologic model WATCLASS, is a coupled land surface scheme–hydrologic model that has been developed by combining the land surface scheme CLASS with the hydrologic model WATFLOOD (Soulis et al., 2000). CLASS is a bulk layer model (Slater et al., 2001), with a separate 1 layer snow cover above the soil, with the ability to consider a partial snow cover, as well as the effects of forest vegetation on snow interception, and the energy balance of snow beneath the canopy. The snow representation of CLASS has been considered previously and was recently tested and compared to a large number of other landsurface schemes by Slater et al. (2001) for a single site in Russia. These tests clearly demonstrated the strengths and weaknesses of CLASS. For example, Slater et al. (2001) found that CLASS tends to consistently hold a snow cover longer, and have a lower ablation rate. Fassnacht and Soulis (2002) considered a number of algorithms aimed at improving the snow components of CLASS. These included the effects of mixed snow/rain at temperatures close to 0^oC, fresh snow density as related to air temperature, canopy snow interception, and maximum snowpack density. The inclusion of these algorithms in CLASS provided improved prediction of these parameters, and had significant effects on heat fluxes. However, they had little impact on streamflow. This is somewhat surprising, and may be an indication of the insensitivity of the runoff routines in WATCLASS to differences in snow regime.

Soil temperatures beneath a snowpack play an important role in controlling the timing and rate of snowmelt. Methods to estimate soil temperature include both heat conduction techniques as well as the force-restore method. The heat conduction method provides robust results, but is reliant on detailed vertical profiles of soil properties and temperature, and are prone to drift over longer periods of time. The FRM is a relatively simple approach developed for estimating diurnal variations in soil temperature. Hirota et al. (2002) suggested a modified FRM for estimating deep soil temperature. The extended FRM was used both in a stand alone mode, and also to determine the bottom boundary layer for a heat conduction model.

Snow Ecology

Jones et al. (2001) provided a complete review and synthesis of snow ecology. This review considers snow, climate, physical properties of snow, snow chemistry, microbial ecology of snow and freshwater ice, effect of snow on animals, snow-vegetation interactions in tundra environments, and the use of tree-ring dating of past snow regimes. An interesting aspect of this work is the attempt to bring all of these aspects of snow together. For example, the editors note the implications of climate change on the snow ecosystems, with a warming climate reducing snow cover and duration, with changes in land-atmosphere energy and water fluxes. These would then modify weather systems and growing seasons, and the emission of greenhouse gases from the soil and snow systems. These changes would in turn have effects on the microbiological activities, and associated invertebrate populations and vegetation. Although Jones et al. (2001) outlined many of the important processes and linkages between the physical and biological systems related to snow, there is obviously a tremendous amount of research required to better understand these linkages, and to apply these to better understand the future effects of climate change.

FROZEN SOILS AND PERMAFROST HYDROLOGY

Organic soils are ubiquitous in the subarctic and the Arctic. Although wetland literature has contributed much to the understanding of the properties and the hydrological behaviour of peaty soils, their hydrological function in the permafrost environment has received only meagre attention . Recently, however, several papers have focused upon a number of aspects on the frost hydrology of organic-mineral layers, on the scale of hillslope but with implications that extends to catchment runoff. Research was carried out primarily in two field areas: one in the arctic where the tundra (Siksik Creek site) is dissected into earth hummock fields, the other in the subarctic (Wolf Creek site) where hillslopes of different orientations and elevations are clad by shrubs and woodlands of divergent species. The results offer illustrative examples of the hydrological responses to different arrangements of the organic and mineral soils.

Stratigraphy and hydrological properties

Many permafrost areas have combinations of organic and mineral soils with different hydrological properties. The organic layer is often highly porous and sometimes with well developed profiles that generally consist of an acrotelm (hydrologically active) and a catotelm (humified and compacted) layer. Quinton et al. (2000) noted that within a depth of 0.3 m, the hydraulic conductivity can change by three orders of magnitude, from almost 10^2 m/d near the surface, to $<10^0$ m/d at depth. Other hydrological properties such as porosity, specific yield and retention, also change with depth. Infiltration into the frozen organic layer is usually unimpeded since the soil pores are seldom fully ice-filled. It has been demonstrated experimentally in Alaska and the Canadian High Arctic that upward vapour flux from the organic soil in the winter involves sublimation of ground ice in the soil, thereby further reducing the ice content to facilitate meltwater infiltration. The mineral soil layer, on the other hand, has hydraulic conductivities much lower than those of the organic. The ice content varies with soil type but freezing changes the perviousness and the water holding capacity of the mineral layer.

Flow mechanisms

In hummocky permafrost terrain where the mineral hummocks are separated by peat-filled inerhummock channels, large contrast in hydraulic conductivity gives rise to preferential flow in the organic materials, mainly in the acrotelm (Quinton and Marsh, 1998). Soil pipes also occur in the acrotelm. These conditions tend to concentrate slope runoff in the interhummock channels, within which flow delivery is particularly effective through matrix flow and pipeflow in the upper peat layer, capable of delivering 0.1 to 1.0 m^3/d per unit width (Quinton and Marsh 1999). Such subsurface flow discharges are as rapid as surface runoff, and subsurface drainage is the predominant mechanism of water transmission (Quinton et al. 2000). Lateral flow is obstructed by the mineral earth hummocks so that the flow paths tend to follow a tortuous network that consists of primary drainage oriented in a downslope direction, fed by a number of secondary channels that collect water across the slope. Increasing tortuosity lengthens the time of runoff delivery along the hillslopes (Quinton and Marsh 1998).

Like the hummocky areas in the tundra, permafrost slopes in the subarctic open woodland have quick and slow flows. On theses slopes, an organic layer of varying thickness drapes over mineral soils. Deep organic soils may be distinguished into acrotelm and catotelm zones, in which case quick flow is restricted to the top layer. Where there is no distinctive catotelm development, the mineral substrate is the only slow flow zone, with considerably lower hydraulic conductivities than the organic soils. This gives rise to a two-layer flow system in which the bulk of runoff is shed through the organic layer as quick flow (Carey and Woo 2001a). The changing positions of the frost table and its overlying saturated zone govern whether quick flow and/or slow flow will occur.

In the spring when the frost table is still shallow, water can infiltrate the highly porous (but frozen) organic layer, but percolation is retarded by the frozen mineral substrate. Rapid meltwater infiltration coupled with much slower downward percolation usually leaves a perched saturated zone above the organic-mineral interface (Carey and Woo 1999) to support matrix flow and pipeflow in the peat layer. Surface runoff along rills may be produced briefly, for one to several days (Leenders and Woo 2002). As ground thaw continues and the frost table drops within the mineral layer, slow flow may be produced but its magnitude is meagre compared with quick flow. Quick flow ceases when the water table falls below the organic layer.

The role of ground ice in the soil, or the absence of it, has been emphasized. Leenders and Woo (2002) examined runoff from the subalpine willow-shrub to the subarctic open woodland downslope, both with a two-layer flow system. They measured slow flow in the ice-poor mineral soil which, though allowing downward percolation, yielded negligible lateral drainage. In a simulation study using the SLURP model, they found that neither increasing the snow water equivalent nor changing the temperature (which affects the calculated melt and evaporation rates) has much effect on the flows but raising the ground ice content in the soil, leading to reduction in percolation, would enhance lateral fast flow from snowmelt.

The role of pipeflow in the frost-affected two-layer system was studied by Carey and Woo (2000). Pipes generally occur at the organic-mineral interface and are usually linked to rills (which may develop when the pipe roofs collapse) to form drainage networks that can convey water rapidly downslope. The Manning=s equation is found to be applicable in representing the flows. The occurrence and magnitude of pipeflow is markedly controlled by the frost table: a shallow frost table in the spring can inhibit deep percolation to enable the saturated zone to remain near the surface, to be tapped by the pipes; whereas a deep thaw during the summer, dry period drops the water table below the organic layer and pipeflow ceases.

Hillslope-streamflow connections

In the arctic tundra, Quinton and Marsh (1999) observed that the peat thickness is about 0.3 m on most of the slopes, but reaches 0.4-0.7 m near the streambanks, apparently separating the hillslope into a near-stream and an upland zone, the latter covering almost 90 percent of the study basin. During the wet period, the extent of the saturated area reaches most parts of the slopes. With a high water table, slope runoff is conducted principally through the upper peat layers while seepage from the catotelm is limited. Then, hillslope-streamflow connection is facilitated by (1) flow contribution from large distances upslope and (2) a high water table that permits rapid flow delivery in the acrotelm. The duration of high flow contribution to streamflow varies, depending on the width of the near-stream saturated area, gradient and length of the slopes, the snow water equivalent the melt rate (Quinton and Marsh 1999). In the dry summer season, saturation is confined largely within the near-stream zone and the upland is then hydrologically disconnected from the streams. An expansion of the runoff source area by summer storms will be delayed until the storage capacity of the near-stream zone is satisfied.

In the subarctic with discontinuous permafrost, Carey and Woo (2001b) demonstrated that not all hillslopes yield runoff to basin discharge. Four slopes with distinct topography, soil, frost and vegetation were studied. A south-slope with seasonal frost in ice-poor silt permits deep percolation at the expense of lateral flow, and such a slope makes no contribution to streamflow. Only slopes with a two-layer flow system and with much ice in the mineral substrate yield lateral fluxes in support of stream discharge. Snow usually accounts for about half of the annual water input and the melt period is when most slope runoff is delivered. In this period, slopes with the greatest snow water equivalent and thin organic layer generate more runoff than the other slopes, mainly due to more water being available and less storage capacity to withhold the meltwater. The lower slopes may receive lateral inflow from the higher elevations and this can maintain discharge for an extended period. Where such inflow occurs, the recession limb of the streamflow hydrograph shows two segments that reflect different source areas of stormflow (Carey and Woo 2001a). A steep segment represents the local (lower slope) contribution and a gentler segment indicates inflow from upslope. The extent of the flow contributing area is variable, being much controlled by the hillslope wetness and the properties of the organic layer. In summer, evaportranspiration often exceeds rainfall (Carey and Woo 2001b) and only slope segments with continued inflow would sustain runoff to enhance streamflow discharge.

General hydrological characteristics

The general features of organic soils in a frozen ground environment may be summarized as follows.

(1) Infiltration into the frozen but highly porous organic layer is not limited

(2) Subsurface flow is the principal mode of runoff, with most of the flow passing through the organic soil regardless of whether it lies on the mineral substrate or infills the channels between the mineral earth hummocks.

(3) Quick flow and slow flow can be distinguished: in a two-layer system with peat overlying mineral soil, quickflow occurs in the porous organic cover while in well developed peat accumulations, quickflow is in the top acrotelm layer.

(4) Pipeflow and matrix flow are the major mechanisms of quick flow delivery.

(5) Where the soil is not ice-rich, percolation allows deep drainage whereas an ice-rich mineral soil will prevent percolation.

(6) The frost table in ice-rich soils restricts downward percolation and facilitates the presence of a saturated zone above it so that if the frost table is shallow, the saturated layer would reside in the organic soil to enable lateral discharge.

(7) Spring snowmelt is the period of maximum slope runoff and low summer runoff (which may be absent) is sustained slow flow in the catotelm or the mineral layer.
(8) Thee source area for streamflow is highly variable in space and in time, and is dependent not only on water supply and slope wetness, but also on the properties of the organic layer (such as thickness, porosity, piping).

The flow systems
Different arrangements of organic and mineral soils will affect soil runoff processes differently. Several cases become apparent from the results of these studies.
(1) no organic soil, mineral soil with low ice content (e.g. south-facing slope in Wolf Creek):
vertical processes dominate and runoff is negligible.
(2) the active layer lies entirely within the organic soil (e.g. west-facing slope in Wolf Creek):
high ice content near the permafrost table prevents percolation into the frozen zone below; the soil may develop distinct acrotelm with quickflow and catotelm that only support slow flow.
(3) organic soil between earth hummocks (e.g. Siksik slope):
flow and storage are focused within the interhummock zone with organic soil which may have an acrotelm and a catotelm.
(4) organic above permeable soil with low ice content (e.g. shrubland in Wolf Creek):
easy infiltration and percolation, with flow increasing when the saturated zone rises from the bottom.
(5) organic above soil with low permeability (e.g. north-facing slope in Wolf Creek):
a perched water table is formed during the melt season at the organic-mineral soil interface; a distinct two-layer flow system is developed to include quick flow (in organic) and slow flow (in mineral layer).

Subarctic Canadian shield
The Canadian shield in the subarctic is a mosaic of Precambrian bedrock uplands separated by soil-filled valleys that also contain lakes and wetlands. Discontinuous permafrost underlie the rolling terrain. Contrary to popular belief, the crystalline bedrock outcrops are not necessarily impervious to water entry because there are many rock fractures and cracks to provide avenues for meltwater and rainwater seepage (Spence and Woo 2002). Furthermore, sporadic presence of thin soil patches on the rock can absorb water input and gradually release the moisture to bedrock infiltration or to evaporation. Thus, the runoff ratio from the shield upland is highly variable, depending on the density and size of the bedrock cracks, the patchy soil cover and the intensity of water input. Frost has no apparent effect on the hydrological behaviour of the bedrock uplands as the cracks are seldom ice-filled to the extent that would inhibit infiltration in the snowmelt period.

CONCLUSION
This review of Canadian research in snow, frozen soils and permafrost hydrology has identified advancements over the previous 4 years. This has continued to include advances in process studies, model development, and snow ecology for example. Continued advancement of appropriate hydrologic models, and the linkages between atmospheric and hydrologic models has resulted in a suite of models that are available for both use in scientific studies of the interactions between climate, snow and permafrost hydrology, and which will soon be available for use by the use community. However, much still needs to be done in order to better understand the linkages between climate, snow, hydrology and ecology.

Advances in river ice hydrology

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Abstract:
We identify three significant recent advances in the hydrology of river ice: (1) by bringing together disparate information, excellent review articles (Shen 2003; Prowse 2001a&b; Beltaos 2000) have noticeably advanced our appreciation of river ice hydrology. Recently, there have also been two special journal issues on river ice (CJCE 2003 and HP 2002) and five conferences (CRIPE 1999, 2001 and 2003 and IAHR 2000 and 2002); (2) There has been the recognition that in order to advance, collaborate efforts are required (Prowse, 2001b) and Shen (2003) has specifically called for a collaborative research project; and (3) perhaps the greatest advance is the fact that there has been the birth of the discipline of the *hydrology* of river ice: The case has been made that river ice is too important to ignore when studying water quantity (Water Survey of Canada has launched a program to address this issue), water quality (temperature, dissolved oxygen, nutrients and pollutants), sediment transport and geomorphology (particularly as it relates to breakup), stream ecology (plants, food cycle, etc.) and fish habitat, behaviour and survival. There have also been significant advances in modelling (1-D public domain ice jam models are now available; the first public-domain 2-D model capable of simulating flows with an ice cover is now available and a commercial version of a complete 2-D ice-process model is being completed). The main need for further work is to: (1) interface geomorpological and habitat models with these river ice hydrodynamic models; (2) develop a complete package (database management, remote sensing, forecasting, intervention methodologies, etc.) to better intervene in ice jam induce flash floods (e.g. Badger Nfld, February 2003). We would add that, given the importance of winter navigation on the St. Lawrence River to the Canadian economy (approximately $2 billion annually), a durable and dependable solution to prevent ice jams downstream of Montreal is still required.

INTRODUCTION

On one level, the need to include river ice processes in the study of hydrology in geographical areas subject to freezing temperatures seems ridiculously self-evident. For example, in Canada, our rivers and lakes are frozen a substantial portion of the year. To ignore ice and its impact on the water cycle seems therefore completely naive. However, it is amazing that the winter period is still usually ignored despite the fact that the issue of 'riparian flow' (the minimum flow release that flow regulators and water users must guarantee to support the aquatic environment) is currently a very critical concern. In April 2003, there was a major workshop on this issue in Quebec City that brought together most of the key provincial players (some of the given conference papers will eventually be published). When one of participants was asked if river ice issues were addressed ('by at least one of the speakers'), he said that the workshop did not address them because river ice processes have no impact on fish (which in fact contradicts the findings of all researchers exploring this issue to date, (e.g. Brown et al.1999)

At a recent workshop in Quebec, Hamilton (2003) provides another example of the disregard for river ice. He suggested that the "what we don't see can't be important" attitude of some winter-hydrological-information-providers is in total contradiction to the map-makers of the 18th century who surrounded their regions of 'no data' with scary dragons and monsters to indicate to the navigator that the uncharted waters hold many potentially very dangerous surprises. He then made the case that at many gauged locations, the most extreme events (low and high stages) are related to river ice processes and that in ignoring the winter period, we may be ignoring the most ecologically important period (see Table 1).

THE BIRTH OF RIVER ICE HYDROLOGY

Despite this disregard by some hydrological researchers of river ice processes, we believe that during the last few years, we have witnessed the birth of the discipline of the *hydrology of river ice*. Our hope is that this article lends further credence to this assertion. That being said, we recognize that some may argue that this statement is a gratuitous exaggeration for the following reasons.

> There have been annual conferences on river ice for decades and, after all, Ashton (1986) already presented the hydrology of river ice as a formal subject in the first chapter of his book. As well, articles on the ecology of river have been around for at least 10 years (e.g. Gray and Prowse 1993; Prowse 1994). In fact, there is even a review of the hydrology of river ice given by Beltaos in 2000. Furthermore, the hydrology of river ice has always been an implicit part of the engineering-related studies of river ice that date back many decades.
>
> Others may also argue against the assertion for a totally different reason. They may say that the scientific community does not yet recognize the importance of the subject since river ice issues are discussed in specialized and marginal articles only; those who do 'core' hydrology have no need for river ice nor do they make reference to it.

Finally, some would also say that its socio-economic impact is negligible since it seems that every publication related to river ice still has to start with an explanation of its relevance.

Perhaps these groups are right. Time will tell. However, we believe that we have indeed witnessed the birth of this discipline in the last few years:

To the first group we would suggest that whereas there has always been some hydrological river ice activity, most of the traditional research was primarily aimed at solving engineering problems. It is only recently that studies of a specifically hydrological nature have being carried out.

To the second group, we suggest that the importance of river ice on streamflow, on sediment transport, on pollution, on geomorphology, on the fauna, on fish habitat and on fish health can no longer be overlooked (e.g. see Prowse and Culp 2003; Prowse and Ferrick 2002). We would further suggest that the link to climate and climate change has also been established (Beltaos 2001; Beltaos 2002). We will present this data more fully later on.

To the third group, we would suggest that the social-economic impact is significant (Beltaos 1995; Burrell and Holder 2003). *'The breakup of ice in Canadian rivers and the ensuing ice jams have a multitude of socio-economic impacts such as flooding, damage to private property and infrastructure, interference with navigation, and inhibition of hydropower generation. The total tangible annual cost of ice jams to the Canadian economy has been estimated as $60 million (Gerard and Davar 1995), comparable to the $100 million estimated for the United States (Carlson et al. 1989). A much greater amount is attributed to missed hydroelectricity production opportunities because of inadequate understanding of river-ice processes in general (Raban 1995). In New Brunswick, where detailed damage records are available, it has been found that ice jams cause a third of all flood events, but appear to be more destructive than open-water floods because they are responsible for two-thirds of all flood damage (Humes and Dublin 1988).'* (Beltaos 2002).

To this we would add that the 1993 ice jam in the St. Lawrence river stopped commercial navigation during a 40-day period (Morse 2001). Given that the Port of Montreal by itself grosses $1200 million annually, the impact of this one jam on transporters and clients could easily be $200 million.

In fact, river ice impacts are felt all around the world: "*A significant number of the rivers in the Northern Hemisphere are annually affected by river ice. A river ice season (freeze-over to break-up) of more than 100 days characterizes most of Scandinavia, Canada and Russia, and as far south as 42 and 30 degrees N in North America and Asia, respectively…. Despite the large spatial and temporal extent of this region, little is known about the large-scale hydrologic importance of river ice, except that gleaned from specific case studies. This is surprising given that the results of such studies indicate that in-channel processes associated with river-ice effects are often more important than those on the landscape in determining hydrologic extremes….*" (Prowse and Beltaos 2002).

We believe that recent concerted efforts to bring the hydrological, climatological and social issues to the scientific community in these formal reviews and presentations marks the beginning of the discipline in its own right. Its birth coincides with the increasing awareness of the (a) dearth of winter data; (b) the importance of an ecological approach to ensure sustainable living and development; (c) the importance of extreme events on the health of ecosystems; (d) the concern over climate change; and (e) and the importance to develop strategies to prepare our responses to changes in the environment. We believe that the case has now been made that river ice significantly impacts all aspects of hydrology and northern ecology. Therefore, there will be consequences for researchers in the traditional hydrological disciplines working in cold climates who ignore its significance.

In the following sections, we present the key references to river ice engineering and hydrology. We then present recent advances regarding specific issues. Finally, we suggest future directions.

BACKGROUND: KEY REFERENCES TO RIVER ICE ENGINEERING

The fundamental problems of ice engineering are primarily related to: floods; navigation through ice-infested waters; hydroelectric installations; transportation over ice and offshore petroleum exploration. The following provides some key references that present river ice engineering: Michel presented the ice regime of rivers and lakes (1971) and ice mechanics (1978). Since that time, Canada has played a key role in river ice research. The Canadian Committee on River Ice Processes in the Environment (CRIPE) began to hold its biannual workshops in 1980. They have all been compiled on 4 CDs and are distributed on behalf of CRIPE by the second author. CRIPE originally started under a National Research Council banner but is now under the Canadian Geophysical Union – Hydrology Section (CGU-HS).

CRIPE has also been instrumental in publishing a comprehensive work on river ice jams (Beltaos 1995) and a primer on the hydraulics of river ice (Davar et al. 1996). A revised version will soon be available. CRIPE, in partnership with the Canadian Society of Civil Engineers, is also in the process of publishing a book on breakup. Canadian leadership has also been provided by the National Research Council particularly related to forces on structures (Johnston and Timco 1999; Timco 2001; Fredreking 2002).

Another centre of river ice research is of course the Cold Regions Research and Engineering Laboratory (CRREL) that has produced a number of publications on all aspects of subject matter (e.g., Tuthill and CRREL, 1999; http://www.usace.army.mil/inet/usace-docs/eng-manuals/em.htm).

Ashton (1986) presents a comprehensive book on river and lake ice engineering, he presents a brief overview of the hydrology of ice, the history of ice research, points to the importance of the international collaborations on the topics (a point supported by Prowse and Ferrick 2002) including the International Association for Hydrological Sciences (IAHS); the International Association of Hydraulic Research (IAHR); the Permanent International Association of Navigation Congresses (PIANC) and the Port and Ocean Engineering under Arctic Conditions (POAC). Since 1969, these organisations hold (normally biannual) workshops and/or conferences that are still very vibrant today.

KEY REFERENCES TO RIVER ICE HYDROLOGY

River ice hydrology was formerly an implicit an integral part of river ice engineering and as such all the references quoted previously also form the foundation of research into the specifically hydrological aspects of river ice. Although there will always be some overlap, it is only recently that the science portion is extracting itself from engineering dominated issues. The Canadian leaders in this effort are unquestionably Prowse and Beltaos.

Beltaos, as chairman of CRIPE for the past 12 years, encouraged it to keep an emphasis on the environmental aspects of river ice; he published articles on the effects of climate in 1997; wrote the overview of advances in river ice hydrology for Hydrological Processes (HP) in 2000 and continues to present work on the subject as, for example, in the special issue of HP on the matter in 2002.

Prowse et al. initially edited the Environmental Aspects of River Ice in 1993; presented the significance of ice to streamflow in 1994; authored articles on river ice ecology in 2000, 2001a&b and authored or co-authored four other articles in that special issue of HP in 2002. His 2003 article with Culp in CJCE is particularly thorough in making the link between breakup and other key hydrological issues including all key abiotic and biotic processes.

RECENT ADVANCES IN STREAMFLOW AND RIVER ICE

Across Canada, water quality issues are becoming a critical concern in winter, since this is typically when the lowest flows occur, and therefore when effluent dilution capacity and oxygen replenishment are at a minimum. Increasing pressures on water quantity and quality, in response to economic development, have resulted in a need to be able to accurately quantify river discharge throughout the entire year, rather than just in the open water season (Suzuki et al. 2002, Hamilton et al. 2001; Wang 1999). Despite this urgent need to be able to quantify discharge during low flow, the Water Survey of Canada (WSC), the division of Environment Canada responsible for measuring river streamflow, currently has no means of obtaining accurate winter discharge measurements on a real time basis.

Determining the discharge at streamflow gauging sites is a relatively straightforward procedure under open water conditions. Data collected with automated water level recorders can be readily converted to discharge using established stage-discharge relationships (i.e. rating curves). Ice conditions, specifically ice thickness and underside roughness, are influenced both by flow conditions and by the prevailing meteorological conditions and therefore there is no unique relationship between stage and discharge. The problem is particularly severe during freeze-up and breakup, since discontinuous ice coverage, and dynamic formation and ice clearing processes result in complex flow physics. As an ice cover forms on a river, water is abstracted from the river to form the ice sheet. All the while, the formation of the ice sheet causes an increase in resistance to flow. The stage rises and therefore even more water goes into storage. While water goes into storage, the flowrate downstream diminishes.

Based on analyses of Water Survey of Canada's (WSC) data, Moore et al. (2002) state that: *"Evidence was found for an abrupt decrease in discharge at freeze-up in one of the case studies and for 10 of the 25 stations in the synoptic analysis that had measurements within 30 days of freeze-up (an additional 12 stations had no measurements within 30 days of freeze-up). However, given the paucity of measurements in the early winter, the magnitude, duration and frequency of these events cannot be specified."* To simulate the phenomenon, they built a storage-depletion model that represents streamflow as outflow from two parallel linear reservoirs.

At breakup, ice melts and (or) flows downstream. When this is combined with the decrease in resistance due to the breakup of the river's cover, the flow rate can increase substantially. Prowse and Carter (2002) explain that, based on a case study of the Mackenzie River, water released from hydraulic storage due to flow abstraction during the early winter is shown to be a major component of the flow volume that comprises the following spring freshet. The amount of water placed into hydraulic and ice-growth storage over a 60-day period was calculated to be equal to 27% of the flow that would normally have occurred during this period if an ice cover had not formed. The amount of water released from this ice-induced hydraulic storage at the time of break-up and the concomitant spring snowmelt peak accounted for

15 to 19% of the spring freshet volume, depending on the temporal definition of the 'release' period. The ice-related contribution increases to as much as 25% of freshet volume if the ice growth during the fall depression is also included. On the other hand, interpretation of water level fluctuations at gauging sites using a conventional rating curve approach can lead to overestimation of discharge. Healy and Hicks (2000), investigating the Athabasca River, AB, found that during the early breakup period minor ice movement in the vicinity of gauging stations can lead to errors of up to 300% in the published discharge data.

Hamilton (2003) argued for the need to better gauge and study the impact of ice on the determination of streamflow. He states that in Canada, 18.7% of WSC station days are influenced by ice (Table 1). Bourdages of WSC has established a project to address this need, organising a workshop (2003) on the subject where the use of new acoustic and radar technology was discussed; the use of hydraulic models was presented; the creation of a database presented and a tool for quality control to estimate the impact of ice was presented. The use statistical models was also presented earlier (Ouarda 2001) and this year, Ouarda led a workshop in Quebec City (e.g., Hamilton, 2003) on the subject where the use of neural networks and satellite technology was discussed.

Table 1. Ice affected discharge data in Canada based on WSC data (after Hamilton 2003).

province	Station-days	'B'-days	% of record	% of low extremes	% of high extremes
BC	8 627 563	1 225 202	14.2%	34.6%	0.1%
ON	7 105 110	1 041 975	14.7%	11.9%	7.2%
AB	5 037 259	1 068 452	21.2%	55.9%	11.4%
PQ	4 895 833	890 388	18.2%	26.3%	1.2%
SK	3 320 473	545 186	16.4%	50.0%	26.7%
MB	2 664 249	672 512	25.2%	64.6%	26.5%
NF	1 042 414	192 465	18.5%	25.0%	4.4%
NB	1 022 985	255 992	25.0%	23.6%	7.4%
NS	836 907	86 935	10.4%	1.9%	3.6%
NT	680 209	293 933	43.2%	75.8%	17.0%
YT	666 579	341 715	51.3%	92.6%	2.2%
NU	269 753	150 314	55.7%	89.8%	11.4%
PEI	140 970	13 677	9.7%	10.2%	10.8%
total	**36 310 304**	**6 778 746**	**18.7%**	**38.0%**	**9.3%**

RIVER ICE AND EXTREME EVENTS

Beltaos (2000) points out that the impact of river ice on streamflow and stage is even more severe for extreme events than for average flow conditions. Based, once again, on WSC data, Hamilton (2003) demonstrated that the extreme low discharge is affected by ice 38% of the time and maximum flows are affected 9.3% of the time. (Note that in the case of stage, the impact on flood levels may be even greater. For example, one-third of the floods in eastern Canada are related to ice jams (Beltaos 1995)).

Presently, there are important initiatives to better acquire and manage streamflow data in Canada (Liu 2003), the USA and by globally-based organisations. Combined with new instrumentation technology, new database construction, new international protocols, better quality control methods, new internet capability and the growing recognition that reliable winter data is important, we hope that over the next 5 years, better river ice affected information on discharge and stage will be available.

BREAKUP

Using photos, figures and sketches, Jasek (2003) wrote an excellent article describing dynamic breakup events. Based on field data, he explains the dynamic nature of the process referring to a historically documented case on the Athabasca River, AB where the river rose 17m in one hour and surges traveled at 5m/s. Using a non-dimensional parameter defined as the ratio of the distance downstream from the point of jam release to the length of the jam, he quantifies surge speed and attenuation for both water and ice. Elsewhere, Guo (2002) presents criteria for the onset of breakup.

Prowse and Culp (2003) present the impact of breakup on river ecology. Here are some excepts:

- *'River-ice breakup was shown to be a major, if not predominant, source of floods on cold-regions rivers.*

- *'A flood can produce a major, even temporarily traumatic, reset of ecological processes and forms, biotic and abiotic… More generally, as floods have been identified as primary controllers of many ecosystem-level processes, they have also been gradually incorporated into broad theories of river ecology… notably… the "river continuum concept" (RCC) and the "flood pulse concept" (FPC).* (They conclude ecological studies using RCC and FPC should make the link with river ice.)
- *'Many studies that indicate that breakup floods have far lower recurrence intervals for at least the large-order events.*
- *'The high flow velocities, high stage, and mechanical action of ice combine to make breakup a highly erosive agent of change for river channels, bands, and adjacent riparian zones.*
- *'Sediment concentrations can reach >1000 mg/l during breakup surges… several times greater than the open-water values for an equivalent discharge.*
- *'Breakup can reverse the erosive process in bends…; river deltas are prone to channel shifting…; breakup erosion of banks has even been hypothesized as being responsible for high-level benches;… ice breakup plays a major role in channel enlargement but this remains a point of debate…*
- *'High gradients of water temperature may exist at breakup…*
- *'Breakup … also affects the riparian and aquatic vegetation communities found in and along the channels, including extensive floodplain areas. The floristic composition of such vegetation is influenced strongly by the frequency and severity of the ice and flow actions that characterize breakup.*
- *'Breakup can be considered a disturbance event that can threaten all winter live-stages in ice-covered lotic systems, the less mobile stages being the most susceptible to damage.*
- *'Breakup is known to affect a number of water-quality parameters, ranging from temperature to nutrients, and to even affect the movement of sediment-related contaminants.*

FORECASTING ICE JAM FLOODS

Ice jam floods occur for a number of reasons. Ice jam formation (either due to natural causes or hydro-power peaking) causes an obstruction to flow, which results in high water levels and the potential for flooding adjacent to and upstream of the ice jam. If the ice jam suddenly releases, the resulting high velocity surge of water and ice can lead to significant increases in water level which pose a risk to human life and property downstream of the ice jam location. For example, "on February 15, the more than 1,000 residents of Badger, Newfoundland were forced to evacuate from their homes when a massive ice jam sparked major flooding from three rivers, leaving their town rapidly encased in ice and water" (http://www.redcross.ca/article.asp?id=002379&tid=001). This event is a reminder that there is still a great need to improve our capability to forecast ice jams and the associated flood levels.

One factor which has significantly limited our ability to forecast ice jam formation and release events is the fact that few have actually been documented scientifically. This can be attributed to both logistical and safety reasons. Because of our current inability to predict their occurrence, it is difficult to be in the right place at the right time to facilitate scientific monitoring efforts. In addition, it is usually too dangerous to even attempt to the most basic hydrodynamic variables (ice thickness, flow discharge, and streamflow velocity) during ice jam formation and release events. Consequently, some researchers have used experimental studies to investigate the dynamic aspects of ice jam formation (Zufelt and Ettema 2000, Healy and Hicks, 2001).

Based on field data, Jasek (1999, 2003) documents and analyses ice jam release events, particularly in terms of surge propagation velocity and its impact on surface concentrations of ice (Khan et al. 2000 explored this experimentally). Kowalczyk and Hicks (2003) present detailed documentation of the wave speed and peak magnitude attenuation of an ice jam release surge at 7 stations over 40 km reach of the Athabasca River, AB, which occurred in 2002.

Because of the complexity of processes involved in ice jam formation, and our lack of quantitative data describing these processes, a purely deterministic approach to the problem of forecasting ice jam occurrence is not yet practical. Consequently, numerous empirical approaches have been explored for predicting the risk of ice jam formation. They include threshold methods, which explore the possibility of some critical value of a hydrometeorological factor (e.g. freeze-up level, breakup stage, cumulative heat input, discharge, etc.) beyond which an ice jam event tends to occur. Threshold models have proven to be of limited success, except in cases where multiple threshold conditions have been considered integrally. Other researchers have employed regression methods, particularly multiple regression techniques, in an attempt to rationalize a problem involving a multitude of complex variables.

White (2003) provides an excellent review of many of these ice jam prediction methods and concludes that *"The complex physical processes involved in the formation of breakup ice jams make them difficult to predict, but the sudden nature of occurrence and high stages make jam prediction desirable. The lack of a complete analytical model describing the processes that lead to ice cover breakup, transport, and jamming prevents the development of general, mechanistically based prediction models. In addition, the lack of an analytical model of breakup jam often results in the arbitrary selection of model variables. The relatively small number of ice jam prediction models that have been* *developed include empirical, statistical, and artificial intelligence techniques. Existing empirical jam prediction methods are highly site specific and can have a high rate of false positive predictions, leading to low confidence and poor response to ice jam warnings… The poor performance of several existing models reinforces the use of model variables that are easily measured, estimated, or forecast to enhance the utility of a breakup jam predication model. Artificial neural network systems appear to hold much promise for the future but require further testing."*

Massie et al. (2002) explored the viability of neural networks, black-box models that can be 'trained' to properly represent complex non-linear cause-effect relationships. Neural networks consist of a set of nonlinear computer algorithms that are "trained" to evaluate an input vector and calculate an output that minimizes error. McDonald et al. (2002) present a successful application for predicting ice jams at a confluence. Mahabir et al. (2002) used fuzzy logic to quite successfully forecast the risk of ice jams for Fort McMurray.

Daly (2002) used a different kind of modelling approach that incorporates real-time observations for forecasting purposes. His results indicate that this method is very promising.

The real-time use of Radarsat and other remote sensing techniques are being studied (Gauthier et al. 2001; Robichaud and Hicks 2001; Weber et al. 2003; Tracy and Daly 2003). Images taken of the 2003 (Pelletier et al. 2003) Athabasca River breakup indicate that RADARSAT may be a very promising technology for this type of river system (large) with this type of RADARSAT coverage (daily as opposed to the 2.5 day interval at other Canadian locals).

RIVER ICE AND GEOMORPHOLOGY

Prowse (2001a) specifically demonstrates the linkages between river ice, sediment transport and water quality. He includes a comprehensive reference list, the most complete ever assembled on this subject. He states that ice can *"significantly modify a number of … geomorphic and chemical processes that have important biological implications, such as the erosion and deposition of sediment or the production and transport of oxygen. Geomorphologic effects are most pronounced at breakup when ice is capable of producing unique erosional and depositional features."* A couple of years later, Prowse and Culp (2003) provide a key overview of the linkages between river ice and geomorphology.

Elsewhere, Milburn and Prowse (2002) discuss the transport of cohesive sediments during the initial phases of breakup; Beltaos and Burrell (1998) present sediment transport during breakup; Smith (2003) presents the findings of flood frequency analysis of the Peace-Athabasca-Delta; Hicks (1993) demonstrates that ice can be the principal geomorphological agent in an anastomosing river system; and Sui et al. (2000) present sediment dynamics in a frazil jammed reach.

However, Ettema (1999a&b) is probably the most active researcher in this area. He states (2002): *"The extent to which alluvial channels respond to ice-cover formation, presence, and breakup is not well understood. Some responses are well known and observed, such as increased flow stage or localized scour beneath the toe of an ice jam. Other responses are known in concept, such as altered bedform geometry, but are not well documented. Some potential responses are barely recognized, such as channel-thalweg adjustment. Many responses are temporal, such as the channel readjusting itself once ice is gone. A few responses may have a more enduring impact, such as a meander-loop cutoff. Most responses have not been investigated rigorously. The responses affect the full gamut of relationships between flow discharge and stage, macroturbulence structures, sediment-transport and mixing processes, and alluvial-channel stability. Of importance are the relative scales of length and time associated with ice-cover formation, presence and breakup, and a channel's facility to respond to ice."*

RIVER ICE AND WATER QUALITY

The quoted studies on sediment transport and breakup speak also directly to water quality issues because sediments (especially fine sediments) are great transporters of pollutants and nutrients. Based on field data, White and Laible 2002, present a finite element model to explore the relationship between dissolved oxygen and photosynthetically active radiation. During breakup, there are very dynamic processes pertaining to water temperatures and the replenishment of the food chain. Prowse (2001a) provides a review of contributions in the water quality subject area. For example, he shows that *"ice controls water-atmosphere exchanges of oxygen and the receipt of radiation such that it becomes the primarily control of the under-ice oxygen regime. Similarly, it directly influences the mixing of dissolved and suspended substances within the water column and the erosion/transport/deposition of suspended material."*

Sydor and Boutot (1999) simulated the increase in minerals concentration in a controlled USA-Canada boundary water by the growth of river ice. Elsewhere, Singh and Hasnain (2002) carried out a detailed geochemical study of the water of the Garhwal Himalaya catchments with the objective of evaluating the weathering and geochemical processes controlling solute chemistry and sediment transfer in the Ganga headwater. These waters are a very important source for irrigation purposes in India.

RIVER ICE AND AQUATIC ECOSYSTEMS

Numerous researchers have documented the impact of ice on fish and the aquatic ecosystem during the 1990s. Prowse (2001b) provides a detailed review and description of the biological effects for all stages of winter ice. *"Special focus is placed on the role of ice in seasonal movements and avoidance behavior of fish and benthic organisms, the creation of unique in-channel and riparian habitats, the modification of aquatic and floodplain vegetation, and some river-ecology* *theories, including disturbance ecology and flood-pulse theory. Included is a comprehensive reference list, the most complete ever assembled on this subject."* During freeze-up, fish seek particular habitats that will allow them to conserve energy, be safe from predators and be sheltered from frazil ice, anchor ice and super-cooled water. Border ice can be beneficial (reduced flows; increased protection from predators and super-cooled water) whereas frazil ice can irritate the gills and impede the exchange of oxygen to eggs in the bed. Anchor ice can suffocate eggs and once released can transport organisms downstream. Once the river freezes over, water levels can increase upstream providing habitat opportunities but can cause severe low levels downstream causing mortality. *"As the ice grows down into the littoral zone, it can also influence the distribution of certain benthic invertebrates. In general, organisms can either migrate to deeper water or stay in a dormant state (diapause) to be trapped by the growing ice… Spring-fed sites offer ideal winter fish habitat because they preclude the freezing of eggs and support large populations of invertebrates for winter feed."* As water level recede during winter, locally suspended ice sheets can provide access for many aquatic mammals such as muskrat, mink, otter, and beaver.

Other recent references include Dolgopolova (2002) who discusses how the turbulent structure of ice-covered flow impacts on river habitat and Partridge (1999) who discusses the impact of ice on invertebrates in an estuarine setting. Studies in Japan (Nogami et al. 2002) show that *"whereas (in winter) migratory fish, in winter, tend to choose a habitat with reduced flows velocity near the river shore, the sculpin, a benthic fish, prefers one with riffles in the central parts of streams and a lot of loose stone on the river bottom, regardless of the season."* White et al. (2002) present *"the first reported information on microbial communities within riverine frazil deposits… that may be important in wintertime DO processes".*

Of particular significance, Alfredsen and Tesaker (2002) are trying to include winter habitat into their overall assessment tools. They note that *"short periods with adverse winter conditions may be the limiting factor for the fish population. A winter habitat assessment procedure will include a study of fish behaviour in winter conditions, the effects of various ice types on habitat selection and the dynamics of ice breakup on the microscale… Our (their) paper outlines how fish respond to winter conditions and how hydraulic, hydrologic and biological modelling can be applied to describe winter conditions and their impact on physical fish habitat. Some of these processes are currently being incorporated into the Norwegian habitat modelling system. This paper also points out several areas where more research is needed to describe the physical processes and the biological responses to a changing environment."*

They come to the conclusion that it is a very tough job. Of particular interest in their paper is their treatment of the macro-, meso- and micro-scales. Whereas overall climate factors (and hydrological models) are the most important driving forces at the macro-scale (i.e., a stretch of river), most hydraulic models concern themselves with reaches (meso-scale) whereas the habitat issues are primarily centered at the micro-scale. Therefore, before we advance in this area, we must have tools that can seamlessly move from one model at one scale to a different type of model at another scale. This will be quite a challenge.

RIVER ICE AND CLIMATE CHANGE

Prowse et al. (2002), Prowse and Bonsal (2003) and Beltaos (2000; 2002) describe the potential links beween river ice and climate change. Prowse and Beltaos (2002) state the following: *"A brief review of the hydrologic aspects of river ice shows strong climatic links and illustrates the sensitivity of the entire ice regime to changes in climatic conditions. To date, this sensitivity has only partly been documented: the vast majority of related studies have focused on the timing of freeze-up and break-up over the past century, and indicate trends that are consistent with concomitant changes in air temperature. It is only in the past few years that attention has been paid to the more complex, and practically more important, question of what climatic change may do to the frequency and severity of extreme ice jams, floods and low flows. The probable changes to the ice regime of rivers, and associated hydrological processes and impacts, are discussed in the light of current understanding."*

"Although some regional records suggest a general shortening of the ice season has occurred in approximately the last century, the records are not consistent. Furthermore, while reductions in the ice season have been linked with air temperature, other hydroclimatic factors such as changes in precipitation could masque or even override such a relationship. Both winter drought and increases in snowfall could have significant effects, especially in the case of ice-jam floods, where changes in precipitation can influence a number of flow and ice characteristics. Overall, changes in almost any of the major meteorological fluxes are capable of producing significant change in ice conditions, including the nature and timing of freeze-up, ice thickness and break-up severity. In addition to having important hydrologic implications, such changes can produce important geomorphologic, ecological and socio-economic impacts."

RIVER ICE BEARING CAPACITY

Kuryk (2003) expresses great concern about the potential impact of climate change on winter transportation. He states that: *"The winter road network in Manitoba spans a length of 2178 km and services 30 communities (approximately 29000 people). It is extremely important for the shipment of goods, employment of locals and travel between communities. With the certainty of climate change and expected temperature increases of 4-6°C by the end of this century, there is a real threat to the seasonal operation of winter roads. The inevitable climate change from greenhouse* *gas emissions will result in later freeze-ups, earlier spring melts and more frost-free days"*. Beltaos (2001) provides a methodology to calculate short term bearing capacity while van Steenis et al. (2003) present a finite element secondary creep model to calculate ice deformation under long term loading conditions.

ICE FORCES ON STRUCTURES

Considerable advances have been made recently concerning ice forces on structures. Important data has been obtained from the monitoring of the Confederation Bridge (Brown 2000). The Canadian Hydraulics Center under the leadership of Timco has built up a database of ice forces on structures (e.g., Johnston et al. 1999; Timco and Johnston 2002; Frederking 2002) and Carter et al. (1998 and 2001) present a comprehensive model of static ice forces on dams. This is a very current issue; for example, the new Quebec law on dam safety requires the systematic review of dam stability. Under the current code, most of the thousands of small dams do not meet the code regarding their resistance to static ice forces. The same is probably true of most other small dams in Canada.

ICE CONTROL STRUCTURES AND ICE BOOMS

Although this may be considered more an 'engineering' topic than a 'hydrology' topic, the subject of ICS is included here because hydrology is always concerned with flood control and the primary purpose of ICS's is to protect residents from ice-jam related floods. Over the last few years, there have been significant advances in this area: A lot of work has been done on the design of ICS's by the group at CRREL (Lever and Gooch 2001; Lever and Daly 2003; Lever et al. 2000) and the group at Laval University (Delcourt 2002; Francoeur 2002; Morse et al. 2003). Also, the design of ice booms has seen great progress recently (Hopkins and Tuthill 2002, Abdelnour et al. 2002; Tuthill and Gooch 1998 & 1999; Morse 2001a&b).

WINTER NAVIGATION

Given the desire on the St. Lawrence to prevent flooding and to open a winter waterway, a large number of river ice studies were carried out on the river in the 1960s. For example, there were studies to see if nuclear power could keep the Seaway open all year round. Although that particular idea was abandoned other studies led to the construction of bridges, booms and artificial islands to control ice and a modern ice breaker fleet was purchased. Winter navigation and flood protection seemed to be fine until a major congestion ice jam blocked the River in 1993. To improve the situation, a number of studies and actions were carried out regarding (a) ice management (Carter 2003), (b) real-time ice information systems (Dumont 2001; Morse and Crookshank 1998), (c) brash ice characteristics (Morse et al. 2001 and Morse and Hessami 2003a), (d) the congestion jam process (Morin et al. 2000) and the use of neural networks to model congestion jams (Morse and Hessami 2003b). Hopefully these studies have helped reduce the risk of jams but further studies are still required to a secure a dependable winter waterway.

Tuthill (1998) presents navigation through ice on the Mississippi. He (1999) also presents Soo Locks ice problems and possible solutions and Tuthill and CRREL (1999) present flow control measure to manage ice. Hopkins and Daly (1999) present a discrete element model of river ice at navigation structures and Liu et al. 2001 *"report on an ongoing two-dimensional numerical ice-hydraulic model study on the dynamics of ice passage at locks and dams in the Ohio River for a generic navigation project... Simulations are carried out for a variety of lock and dam configurations and operation conditions. These simulations include ice accumulation and ice passage in the upper approach of the lock under different dam gate settings, as well as upper lock gate and guide wall configurations. The capability of high flow air bubblers to deflect and retain ice is also simulated by imposing horizontal water velocity distributions at the selected location."*

RIVER ICE AND ESTUARIES

In addition to the work by Partridge quoted above (2001), at the initiation of Burrell (Morse et al. 1999), there has been some headway made into the understanding of ice processes in estuaries. With the collaboration of Hydro-Québec, the interplay between river ice, hydrodynamics and thermodynamics for a mesotidal estuary on the north shore of the St. Lawrence has been documented (Morse et al. 2003). It was found that the ice is very unsafe, that the temperature of the salt water played a key role and that the neap/spring tidal cycle was a real driving force that dominated ice formation and ice breakup.

FRAZIL ICE

There has been a renewed interest in frazil ice in recent years (Terada et al. 1999; White and Acone 1998; Doering and Morris 2003). Frazil affects the behaviour and survival of fish (Brown et al. 1999; White et al. 1999); it impacts sediment transport (Smedsrud 1999); it causes floods (Asvall 1998); it impedes the efficient operation of hydro-electric facilities (Morse and Quach 2002) and it must be taken into account for the design of municipal intake structures (Ettema et al. 2002, Morse and Trudeau 2003). There has been a particular interest in frazil ice on the Yellow River where very impressive field studies have taken place (e.g., Li and Zhu 1999; Ke et al. 2000).

ICE PROCESSES

In addition to the dynamic ice process studies discussed above, over the last few years, studies have continued on other ice processes as well. Shen (2003) provides an excellent overview on ice formation processes, the ice covered period and breakup. He gives a brief explanation of key processes and refers the reader to the most important key articles on each subject. He discusses skim ice, border ice, frazil ice, anchor ice, ice thickening, ice accumulating, ice transporting, ice thinning, ice breaking and ice jamming and it recommends areas for further research (see 'future directions' below). For those who are new to the field, this article is the ideal introduction.

Andres and Van der Vinne (1999) describe the insulating effect of snow on ice cover thickness. Ettema and Zabalanski (2001) discuss the formation of open leads in an ice cover. Timco and Cornett (1999) quantify the angle of internal friction for ice accumulations (required for numerical model simulations of ice jams). Granular ice (marble ice) has been documented in the lab and in the field (Hammar et al. 2002)

NUMERICAL MODELLING

DYNARICE is a coupled Eulerian-Lagrangian model. The hydrodynamics of the flow are simulated with a finite-element scheme and the ice dynamics are simulated with a Lagrangian discrete-parcel method has already been referred to under the heading 'Winter navigation' and the link with fish habitat modelling has already been presented under the heading 'Aquatic ecosystems'. Another published application presents the jam dynamics of the Upper Niagara River (Lu et al. 1999). Further information is presented by Shen (2002) in his overview of numerical modelling of ice processes. Shen's team is working on producing a commercial version of this model. Once available, this model will be a great asset to all hydrologist seeking to evaluate the interplay between river ice and ecological and geomorphological processes; to flood protection agencies to reduce the risks of ice jams and to engineers to help optimize hydro-electric operations and winter waterways. In Canada, Doyon (2001) collaborated with Shen and produced a Canadian draft version of the model.

Elsewhere in Canada work is still on-going on RIVICE which will lead to a public-domain 1-D general purpose ice process model; CRREL researchers continue to improve their 2-D model (e.g., Daly and Hopkins 2001, 2003); there is now a public-domain version of a 2-D hydrodynamic model that allows the user to model specified ice conditions (www.River2d.ca). Blackburn and Hicks (2003) explored the applicability of a public domain finite element model for ice jam surge release modelling. Model comparison and evaluation has also been undertaken regarding ice jam modelling (Healy and Hicks 1999, Carson et al. 2001, 2003).

FUTURE DIRECTIONS

We have summarized our view of future directions in the abstract. Here are a few quotes from the leading researchers that inspired that view:

> *"Despite the great strides made in recent decades in understanding river ice processes, several aspects of ice breakup, movement, and jamming require further study"* (taken from the preface to the special CSCE river ice issue by Burrell and Holder 2003).

> *"River ice has been shown to affect numerous environmental processes and to create unique ecological conditions at all scales of riverine systems, from brooks to major rivers and deltas. Incorporating information about river ice into the ecological understanding of lotic systems, however, will be most difficult for the larger scales. This is primarily because there is no clear theoretical basis for the functioning of large-river ecology even under open-water conditions…. Hence, achieving a greater understanding of the environmental aspects of river ice will not only require further advances in the physical sciences but in the field of aquatic ecology as well. The most efficient method to achieve this is to conduct, whenever possible, multidisciplinary studies of river ice. From a more general river management perspective, it would also seem prudent to begin expending resources to a degree that more closely reflects the proportion of the year that rivers are affected by ice. For many hydrologic regimes of the world, streams and rivers are ice covered for a majority of the year, yet minimal research is conducted during this period compared with the more ''researcher friendly'' open-water period. Without a doubt, scientific progress is hampered by the logistical difficulties and high costs associated with conducting ''winter'' research. Such obstacles, however, can no longer be deterrents if a comprehensive understanding of lotic systems is to be achieved"* (Prowse 2001b).

> *"Areas that are in most need of advancement are suggested. These areas include the development of surface and suspended ice discharges during the freeze up ice run, and the phenomenon of dynamic break up. It is recommended that future research should focus on these weak links in our knowledge in order to provide a comprehensive understanding on river ice processes… it is clear that we still need to develop a concise method to predict the occurrence of ice cover breakup. In particular, the rate and magnitude of the rise of discharge that can trigger a dynamic break up on a controlled or uncontrolled river are to be determined. A joint effort from the ice engineering community might be needed. For example, a coordinated monitoring program on breakup conditions during the peaking of hydropower facilities, could lead to useful database for developing*

these criteria… These studies should make full utilization of available research tools, including field, laboratory, and analytical and numerical methods. Well-coordinated joint efforts with close cooperation of researchers with complementary expertise should be promoted" (Shen 2003).

CONCLUSIONS

This review of the advances in river ice hydrology for the 1999-2003 period leads to the following conclusions:

- Sometime over this period, river ice hydrology as a discipline has been born. Its birth probably coincides with the publications of 2001 articles by Prowse. In these and other articles, researchers in this area have demonstrated that there are very important links between river ice and river discharge; extreme events (floods and low flows); sediment dynamics, geomorphology; dissolved oxygen; pollutant resuspension and transportation; stream and bank vegetation; nutrient fluxes; invertebrate protection and destruction; fish habitat creation and destruction and fish survival.
- There is still a huge need to further research these links and to do so, it is essential to organize collaborative efforts.
- Recent advances in instrumentation and other technologies (satellite and internet) create some new opportunities.
- There have been important advances in low-cost structural control methods to protect against flooding (booms and 'ice control structures') and to increase navigation lock efficiency although there is still the need to strengthen winter navigation in Canada.
- There have been important on-going developments in deterministic numerical modelling and advances in neural network modelling seem very promising.

A revised Canadian perspective: Progress in Glacier Hydrology.

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Abstract:
Current research into glacier hydrology is occurring at a time when glaciers around the world, particularly those whose hydrological regimes affect populated areas, are shrinking as they go through a state of perpetual negative annual mass balance. Small glaciers alone are likely to contribute 0.5 to 1 mm a^{-1} to global sea-level rise, with associated reductions in local freshwater resources, impacts upon freshwater ecosystems and increased risk of hazard due to outburst floods. Changes to the accumulation regimes of glaciers and ice sheets may be partly responsible, so the measurement and distribution of snowfall in glacierized basins, a topic long represented in non-glacierized basin research, is now beginning to receive more attention than it did before. The is partly aided by the advent of reliable automatic weather stations which can provide data throughout the year. Satellite data continue to be an important information source for summer meltwater estimation, as distributed models, and their need for albedo maps continue to develop. This further entails the need for simplifications to energy balance components, sacrificing point detail so that spatial calculation may proceed more quickly. The understanding of surface meltwater routing through the glacier to produce stream outflow continues to be a stimulating area of research, as demonstrated by activity at the Canadian site, Trapridge Glacier, and Canadian involvement in the European site, Haut Glacier d'Arolla. As the Canadian glacier monitoring programme continues to evolve, effort must be directed toward developing situations where mass balance, accumulation, meltwater generation and flow routing studies can be done together at selected sites.

PERSPECTIVE

The view of global change, from the perspective of glacial hydrology, is that we live in interesting times: the Greenland ice sheet is shrinking, we are not entirely sure about Antarctica and small valley glaciers are shrinking around the world (Haeberli and others 1999). The likelihood that land based ice melt is a significant factor in sea-level rise (Gough 1998) is a cause for concern in a world where so many people live close to coastlines. So too is the effect of glacier shrinkage on local stream flow through populated areas, which depend on the fresh water which the associated rivers and streams can supply (Hopkinson and Young 1998). Nevertheless, a recent work on the effect of land cover change on streamflow in the interior Columbia River Basin (Mathuessen and others 2000) makes no mention of glacier wastage, even though the basin extends into Canada, where the ability to honour the terms of the Columbia River treaty may become significantly compromised as current glacier trends continue.

New concerns are being reported elsewhere, notably the hazards presented by the growth of periglacial and supraglacial lakes, as melt accelerates and the potential for outburst flooding occurs (Benn and others 2001, Haeberli and others 2001). Also, an ecological theme is emerging, as seen by one study, in which the role of lake regulation of glacier-fed streams on salmon productivity is examined (Dorava and Milnear 2000). We also see evidence that glacier hydrology is having an influence on the glacial geomorphology of the Wisconsin ice sheet retreat, wherein landforms originally interpreted in terms of ice flow are now being reconsidered in terms of subglacial water flow (Shoemaker 1999).

As stated in my previous report (Munro 2000), progress in Canadian glacier hydrology must be assessed in context of world scientific activity in this area. While this could be done by focusing upon research based at Canadian sites, the effect would be to ignore the fact that the influence and field interests of Canadian glaciologists extend well beyond the boundaries of Canada. Therefore, the approach taken now, as then, is to report on general research in four principal areas of activity over the past five years and to highlight (boldface) work in which at least one of the authors is currently based in Canada, or was based in Canada at the time of publication. We begin with resource status, then follow with the storage phase, the meltwater phase and routing meltwater to runoff. An assessment and outlook conclude the review.

RESOURCE STATUS

Glaciers receive snow during the winter, thus accumulating water in its solid state until summer, when the rise of temperature to the melting point allows it to convert to its liquid state, thus feeding glacier runoff to local streams and rivers. According to how the winter-summer sequence has gone, the glacier may gain or lose mass in the process, thus effecting an annual net mass balance change, ΔB_n, which Oerlemans and Reichert (2000) express in terms of a seasonal sensitivity characteristic:

$$\Delta B_n = \Sigma_{k=1,12}\{(\delta B/\delta T_k)(T_k - \mathit{T_k}) + (\delta B/\delta[P_k/\mathit{P_k}])(P_k/\mathit{P_k})\} + \Lambda \qquad (1)$$

in which T_k is the temperature of a particular month, k, P_k the precipitation, the italicized terms refer to long-term averages and Λ refers to higher order terms which may become significant for large departures from average conditions.

Thus, according to mass balance sensitivities, $\delta B/\delta T_k$ and $\delta B/\delta[P_k/\mathit{P_k}]$, to respective perturbations in temperature, $T_k - \mathit{T_k}$, and precipitation, $P_k/\mathit{P_k}$, glaciers will grow or shrink, with attendant effects upon sea level and local water supply.

The status of the resource reflects a situation in which ΔB_n has been mostly negative over the past century and is likely to remain so over the foreseeable future, such that contributions to sea level rise from small glaciers and ice sheets are likely to be on a par with thermal expansion of sea water. Some large uncertainties attend the estimates, thus the contrast between Mitrovica and others (2001) and Gregory and Oerlemans (1998), regarding the impact of Antarctica. This contrast may reflect the different perspectives which the authors bring to bear on the problem. Mitrovica and others are attempting to reconcile measured changes in world sea level with ice volume change, raising the interesting point that sea level close to areas of deglaciation will actually fall, while Gregory and Oerlemans are simulating future sea level rise, reflecting a widely held view (Harvey 2000) that ice losses from the Greenland ice sheet may be compensated by gains to the Antarctic ice sheet. Despite the fact that the sea level equivalent of the ice reservoir represented by glaciers other than Greenland and Antarctica appears to be relatively insignificant, ice volume shrinkage due to ΔB_n for small glaciers is a significant contributor to sea level rise. Because such a significant contribution is made from such a small reservoir, it is reasonable to expect massive impacts upon local hydrology as the volume continues to decline.

Recent work on glacier variations

A recent development in the area of glacier variations is the breakdown of major ice sheets in to drainage systems, such that one may view Greenland as a clusters of glaciers, each with its own mass balance behaviour (Zwalley and Giovinetto 2001). Although much needs to be done to get the accounting right, remote sensing results point to shrinkage of the Greenland ice sheet (Thomas 2001), mainly through widespread rapid thinning of outlet glaciers by more than 5 m between 1993 and 1999. This is presaged by the shrinkage of small ice caps in Svalbard, where areal extent has decreased by 18 to 44 percent between 1936 and 1991 (Ziaga 2001). Despite expectations that Antarctica could actually grow in times of climate warming, it too shows reports of retreating coastal glaciers, perhaps due to changes in ice flow rather than to alterations to the accumulation regime (Rignot 2001).

The mass balance sensitivity of small glaciers varies with elevation, the greatest sensitivity to warming being found near the glacier snout (Braithwaite and Zhang 2000), so it is not surprising that most alpine glaciers in the world are retreating. This is certainly the case in Europe, where a persistent cumulative mass balance for Glacier de Sarennes, French Alps has been reported over the past 50 years (Torinesi and Others 2002) and where Chiacciaio del Calderone, the southern most glacier in Europe, has experienced more than 36 m of ice loss since the end of the 'little ice age' (D'Orefice and others 2000). Vincent (2002), in a study of four French glaciers, provides more detail to the retreat sequence, finding two relatively stable periods, 1907-41 and 1954-81, two deficit periods, 1942-53 and 1982-99, the latter deficit period showing a sharp reduction.

High sensitivity at low elevation, with consequent large negative mass balance, is confirmed in the North Cascades of North America, where the most positive ΔB_n were found for glaciers that had significant accumulation areas above 2300 m elevation (Pelto and Riedel 2001). South Cascade Glacier, the most studied in that area, has seen a strongly negative ΔB_n of –0.46 m w.e. over the past 40 years, with Blue Glacier, the more maritime of the two showing a ΔB_n of –0.13 m w.e. (Rasmussen and Conway 2001). A nearly coincident time period shows a ΔB_n of -0.28 m w.e. for the more continental site of McCall Glacier, Alaska (Rabus and Echelmeyer 1998). Peyto Glacier, Canada, shows a ΔB_n of -0.49 m w.e. over the 30 year period of its mass balance record (Demuth and Pietroniro 1999). Beyond the generalization that glacier thinning has been most pronounced at low elevations, it appears that there is no clear geographic pattern and no simple relationship between glacier volume change and terminus position change, at least as far as comparisons between Alaskan and North Cascade glaciers indicate (Sapiano and others 1998).

In the Chilean Patagonia of South America, a region not much reported on hitherto, findings are consistent with world wide trends (Winchester and others 2001). Glacier variations in the Hielos Patagonicos of Chile are such that 46 of 48 glaciers are in retreat, contributing approximately 3.6 percent of estimated sea level rise between 1945 and 1986 (Aniya 1999). Harrison and Winchester (2000), looking at glacier variations for the same area during the twentieth century, note increases to retreat rates since 1940 and surface thinning of at least 30 m since 1980. Spectacular ice loss is reported for small glaciers further north, where Glaciar Chacaltaya, Bolivia, exhibiting a mean ΔB_n of −1 m w.e. since 1980, may become extinct in 15 years if present climatic conditions persist (Ramirez and others 2001).

At the moment, Canada's contributions to world wide monitoring of glacier variations are provided through the Geological Survey of Canada glacier mass balance measurement programmes at Helm and Place Glaciers in the Coast Range of British Columbia, Peyto Glacier in the Rocky Mountains of Alberta and White Glacier in Axel Heiberg Island, Nunavut (Koerner and others 1999). All show negative cumulative ΔB_n over their periods of record, consistent with world wide trends. That of White Glacier, in the High Arctic, is approximately half the magnitude of the other three (Haeberli and others 1999).

Weather and climate connections

As indicated in Eq. (2), ΔB_n is controlled by its sensitivities to temperature and precipitation anomalies, the characteristics of which change from year to year in spatially diverse patterns. Temperature sensitivity is the primary suspected agent of glacier shrinkage in a time of global warming, such as that documented for central Greenland, where 1995-99 air temperature appears to be 2 ^{o}C warmer than it was over the 1951-60 decade (Steffan and Box 2001). Precipitation sensitivity is also important, especially in the colder environments of high elevations and high latitude. Holdsworth and Krouse (2002) have shown, using oxygen isotope analysis, that snow zones at different elevations on Mt. Logan, Yukon, can be related to different precipitation source areas within the cyclonic storms that deliver moisture to the area. Such macro-scale systems also deliver moisture to Antarctica, where the primary components of surface mass balance are precipitation and sublimation (Van Lipzig and Van den Broeke 2002).

Temperature and precipitation anomalies also reinforce ΔB_n trends through albedo modification, such as in Greenland, where satellite information indicates a downward albedo trend from 1981-98 which agrees with observed trends in surface melt and precipitation (Stroeve 2001). In fact, Stroeve associates anomalously high albedo values in southern Greenland with intensification of the North Atlantic Oscillation (NAO), a noted feature of the general circulation. The general circulation is also tied to the ΔB_n behaviour of North American glaciers, where warm-dry cycles of the Pacific Decadal Oscillation (PDO) signify negative □B_n, cool-wet phases positive ΔB_n (Kovanen 2003). Associations between the mass balance behaviour of Place Glacier, British Columbia, and the PDO are also noted by Moore and Demuth (2001), where a post-1976 change from cold phase to warm phase PDO appears to have initiated a persistent, more negative ΔB_n and terminal retreat.

Because change in one region of the atmosphere tends to set off other changes far away, a phenomenon which atmospheric scientists refer to a teleconnection, ΔB_n may respond to seemingly unrelated events, such as the El Nino-Southern Oscillation (ENSO) of the tropical Pacific Ocean. The PDO certainly tends to exhibit ENSO-like behaviour (Bitz and Batisti 1999), while Hodge and others (1998) find an ENSO signal in the winter mass balance variations of South Cascade Glacier, Washington, and those of Wolverine and Gulkana Glaciers in Alaska. Also, as McCabe and others (2000) discovered, in their analysis of Northern Hemisphere glaciers, different aspects of the general circulation control the geography of the ΔB_n response, the glaciers of Europe connecting more strongly change in the Arctic circulation, those of North America to ENSO related behaviour.

Better mass balance estimates

The quality of mass balance data is a matter of concern, given their importance in assessing global change and impacts on local hydrology. The major concerns here involve the adequacy of sampling networks on glacier surfaces, whether the surface areas are representative of other glaciers in the region and whether all the factors which determine ΔB_n are being included.

Braithwaite and Zhang (1999) state that the annual mass balance of Greenland is not known with much accuracy, such that the ice sheet can thicken or thin by several metres over the next 20 to 30 years without showing statistically significant evidence of a non-zero balance under present climatic conditions.

A recent look at sampling density on Columbia Glacier, Washington, and Lemon Creek Glacier, Alaska, by Pelto (2000) indicates that significant improvements to accuracy accrue from quadrupling the sampling density, resulting in 4 points km^{-2} for Lemon Creek, 46 points km^{-2} for Columbia. Each density represents a network of 40 stakes, substantially more than are used on the Canadian Glaciers. The possibility of providing better spatial resolution through remote sensing and high resolution modelling on Haut Glacier d'Arolla is explored by Hubbard and others (2000), who find that it compares well with stake measurements. Also, Adams and others (1998) note that the uncharacteristically small Baby Glacier, Axel Heiberg Island, may correlate well will with the mass balance of the larger, more regionally representative White Glacier, thus serving as a useful regional index of ΔB_n.

Cogley and Adams (1998) stress the importance of accounting for internal accumulation in annual glacier mass balances, as well as size bias in surface area representation, finding smaller average ΔB_n as a result, with less impact upon sea level change. Further to this point, Matsuoka and Naruse (1999) report the ability to distinguish winter accumulation from percolated meltwater in ice cores. Elsberg and others (2001) stress the need to account for changes in surface area over time when assessing the effects of ΔB_n upon local hydrology, though for studies of climate sensitivity alone, they note that area changes may be neglected. Accumulated over time, however, area changes may be sufficient to invoke the non-linear term in eq. (2).

THE STORAGE PHASE

The problem at hand is two-fold: to accurately measure precipitation brought to the glacier basin by weather systems in the first place; secondly, to distribute the measurement over the glacier surface in a manner which correctly estimates the temporal and spatial variability of the winter snowpack, a task that must be verified on the glacier itself. There is no straightforward physical relationship for doing this, but a conceptual relationship illustrates some of the basic considerations in estimating incremental snowpack thickness for any one of j points on the glacier surface, h_j:

$$\Delta h_j = f\ \{ P, Z_j, T_j \} \qquad (2)$$

in which P is the precipitation measurement, Zj the elevation and T_j is a topographic shape characteristic, such as a crest, a hollow, or some other charactistic which causes Δh_j to be uncharacteristically large or small for its elevation. To these are added the mid-winter snowmelt and rain-on-snow events which result in complex snowpack structure.

Spatial variability in snowpack thickness can exist on the micro-scale represented by the summer melt surface, as well as over the scale of topographic crests and hollows and the basin scale of elevation change. Thus, there are scaling issues for glacier snowpack distribution which mirror those of snow distribution generally (Bloschl 1999). The challenge is not so much to estimate a value of 569 mm w.e. for the snowpack in one place as it is to know how it may be derived from precipitation gauge data and how representative it may be of the snowpack in its immediate vicinity. Because alpine precipitation measurements are difficult to obtain, they are important in mapping regional snow distributions across Canada for use in general circulation modelling (Brown and others, 2003). There is not much in the current glacier literature on accurate precipitation gauging. That problem is being addressed in the literature elsewhere, though Yang and others (1999) deal with bias in daily precipitation measurements for Greenland. Work has begun to surface, however, on temporal and spatial variability, as well as on sublimation loss.

Temporal variability of snow accumulation

Beginning with Greenland, McConnell and others (2000) model large temporal variability in snow accumulation which is confirmed by ice core data. Thus, ice sheet elevation could vary by tens of centimeters from year to year solely because of changing accumulation. Subsequently, McConnell and others (2001) indicate a measured snow accumulation increase of 37 percent, or modelled increase of 57 percent between 1995 and 1996. As noted in Davis and others (2001), temporal variability can occur in spatial patterns as well.

Work on valley glaciers has emerged regarding year to year variations, as well as change over the accumulation season, now that on-glacier automatic weather station data are becoming available (Oerlemans 2000). Rasmussen and others (2000), looking at the 2100 – 1500 m differences in winter accumulation on Blue Glacier, from 1914 to 1996, note periods of relatively large differences (1914-35, 1945-75) and periods of relatively small differences (1935-45, 1975-96) which are coonected to sea-level pressure changes in the central North Pacific Ocean. Raben and others (2000), using isotope loadings, identify layers and likely synoptic conditions for snow accumulation on Austre Okstinbreen, Norway, thus identifying various stages in the creation of the winter snowpack. Sharp and others (2002) examine the snowpack chemistry of John Evans Glacier, Ellesmere Island, to trace three layers which identify the seasonal evolution of the snowpack: fall accumulation, metamorphosed to depth hoar; winter accumulation, in which snow is mixed with elements of underlying depth hoar; spring accumulation mixed with underlying snow reworked by the wind.

Spatial variability of the winter snowpack

As noted in Fricker and others (2000), it is not possible to determine the mass balnces of Antarctic drainage basins on the basis of currently available accumulation distributions. However, Nereson and others (2000) show that associations can be made with major topographic features, as in West Antarctica, where accumulation south of the local drainage divide is 40 percent less than average for the area, that north of the divide 15 to 40 percent more. On a smaller scale, though one which is still large in comparison to the scales of valley glaciers, Frezzotti and others (2002) show that accumulation in the lee of megadunes, with wavelengths of 2 to 5 km, is 25 percent of that in non-dune areas in the lee of the dunes, 120 percent to windward of the dunes.

In Greenland, Roe (2002) shows that the most important factor for successful simulation of the ice sheet shape is to note that precipitation increases with altitude, according to the ability of the atmosphere to hold moisture. Here, Van der Veen and others (2001) argue that most variance in accumulation data can be explained by the combined effects of the atmospheric circulation and ice sheet topography, effects which can also be seen on small ice caps in Svalbard (Pinglot and others (2001). Bales and others (2001) use kriging of good quality observations to estimate 30 g cm^{-2} a^{-1} average accumulation over the Greenland ice sheet, but note that there are many areas where accumulation is highly uncertain.

Given the relatively small scale of valley glaciers, it would seem that little more is required than an elevation relationship, either in linear form Sharp and others (2002), or expressed as a power law (Bhutiyani 1999). However, Palli and others (2002) show 40 to 60 percent variation in snow cover over short distances on Nordenskjoldbreen, a Svalbard glacier. Further investigation of snow distribution on glaciers is required, beginning with what has been learned about modeling of snow transport in alpine, non-glacierized terrain (Gauer 2001, Liston and Sturm 1998).

Energy exchange and sublimation loss

Mid-winter melt events are not likely to cause loss from the snowpack because the melt is usually captured, where it remains until the summer ablation season. In Antarctica and Greenland, however, there can be significant losses due to sublimation, amounting to 12 to 23 percent over Greenland (Box and Stefan 2001). Where ice is exposed, such as in the blue ice areas of Antarctica, sublimation losses can be double those from snow because of lower albedo and smaller relative humidity (Bintanja 2000). Antarctic work also assesses sublimation losses from blowing snow to be comparable to those from the snow surface itself (Bintanja and Reijmer 2001, King and others 2001). There is no reference here to the works of Pomeroy, cited previously in Munro (2000), an indication that there needs to be more transfer of Canadian work in this area to other venues.

Sublimation losses from snowpacks and blowing snow are not yet extensively reported for valley glaciers. Interestingly, melt loss from accumulation is a consideration for Tibetan Plateau glaciers, where most of the accumulation occurs during the summer monsoon, when high melt levels occur (Fujita and Ageta 2000). These authors also raise the point that because these glaciers are cold based, a substantial quantity of infiltrated meltwater may be refrozen, to remain in the glacier. This would seem to reinforce the point made by Cogley and Adams (1998), to allow for internal accumulation in mass balance calculations.

THE MELTWATER PHASE

The overall problem in this area is to relate weather variations to meltwater production, M_j, by invoking the energy balance equation at any point, j, on the glacier surface:

$$M_j = K_{\downarrow j}(1 - \alpha_j) + L_{\downarrow j} - L_{\uparrow j} + Q_{Hj} + Q_{Ej} \qquad (3)$$

in which $K_{\downarrow j}$ is the solar short-wave radiation at the glacier surface, α_j the surface albedo, $L_{\downarrow j}$ the surface long-wave radiation gain, $L_{\uparrow j}$ the surface long-wave radiation loss, Q_{Hj} and Q_{Ej} the turbulent heat transfers between air and surface due to sensible heat and water vapour, respectively. The term 'surface' is repeated in the symbol definitions because specific surface characteristics, such as condition of the snowpack and whether its removal exposes clean ice, dirty ice or firn, determine how much solar energy will be absorbed due to surface albedo, and what degree of surface roughness will control the turbulent transfer terms. Basin surface characteristics, such as local relief and its effects upon exposure of the sky, affect the radiation terms. Slope angles and aspects are especially important in estimating $K_{\downarrow j}$, while the assumption of a 0 ^oC melting surface is vital to the determination of $L_{\uparrow j}$, Q_{Hj} and Q_{Ej}.

Conceptually, one may view the melt energy supply terms as being external to the basin, those due to radiation coming from the sky, those due to heat and moisture supplied by areas adjacent to the glacier. As noted in my previous report (Munro 2000), each term presents challenges to researchers. Further challenges are presented by problems of temporal and spatial scaling where, in the interests of reducing computation time for large areas and time periods, the objective is to see how long a time step, or how coarse a spatial resolution can be used before significant errors are introduced (Cline and others 1998).

Point studies of surface melt

Surface energy budget studies are consistent in reporting the dominance of radiation in the melt process (Braithwaite and others 1998) and the importance snow and ice albedo to this aspect of the process (Kayashta and others 1999). Results from Glaciar Zongo, Bolivia, document the change in energy balance components from La Nina to El Nino conditions, where El Nino brings a 50 percent increase to turbulent energy transfer, but a three-fold increase in net radiation (Wagnon and others 2001). The net radiation increase is primarily attributed to albedo reduction, due to less precipitation in an El Nino year. Neuman and Marsh (1998) have noted the importance of local advection to Arctic tundra snowmelt as the cover breaks down into snow patches, a process which has also been examined for a glacierized basin in Svalbard (Hodson and others 1998b). The effect of rainfall on melting is seldom described, except when it leads to extreme events, such as the production of slushflows on Brewster Glacier, New Zealand (Smart and others 2000).

Progress continues in the application and development of turbulent transfer theory, from corrections to instrumental errors (Arck and Scherer 2001), to confirmation of bulk transfer theory over melting ice and snow (Denby and Gruell 2000). Arck and Sherer (2002), however, state that not even the bulk profile method will work in all conditions. Sensing the need for simplification in spatially distributed modelling, researchers are seeking ways to reduce computation time by simplifying turbulent heat flux calculation procedures in katabatic flow over glaciers (Grisgono and Oerlemans 2002, Oerlemans and Grisgono 2002).

The main development over the past five years is the growing presence of automatic weather stations (AWS) in glacier research. This allows year-round measurement of data needed for melt calculations, such as solar radiation and surface albedo in summer (Oerlemans and Knap 1998) and snowpack growth over the winter (Oerlemans and Klok 2002). In so doing, they set the stage for studies of winter to summer snowpack transition, such as we see for snow cover in non-glacierized terrain (Kattelmann and Dozier 1999). They also provide data for testing boundary layer models over large expanses of ice, such as Greenland (Denby and others 2002a,b).

Distributed modelling of the melt

Distributed measurement of surface albedo precedes distributed modelling. In the case of Antarctica, Landsat imagery shows the albedo of blue ice to be 0.60, that of snow 0.78, thus partly explaining higher sublimation rates over ice (Reijmer and others 2001). On glaciers, orbiting radar identifies darker imagery in the accumulation zones of Axel Heiberg Island glaciers to be melt zones (Cogley and others 2001). De Ruyter de Wildt and others (2002) use Advanced

Very High Resolution Radiometry (AVHRR) to determine mean albedo over Vatnajokull, Iceland, to assess the extent of the summer ice melt and find that it relates linearly to the specific mass balance. The suppression of albedo from pooling of meltwater in lower elevation zones of Greenland is shown by Gruell (2000), also using AVHRR. According to Gruell, this implies a positive feedback between albedo and melt, accounting for 40 percent of the interannual mass-balance variation wherever meltwater accumulation is important.

Albedo modelling may be done on the basis of measurement data, either by assignment of fixed values to specific cover classes, or by parameterizing according to elevation. Brock and others (2000), working on Haut Glacier d'Arolla, find that elevation parameterization provides little improvement over assuming a constant albedo value over ice. A full set of energy exchange components is needed for a distributed glacier melt model, as demonstrated by Arnold and others (1998) for Haut Glacier d'Arolla, where this is linked to surface routing and subglacial flow models. A spatial modelling approach is used for a sub-catchment of Haut Glacier d'Arolla by Willis and others (2002) to show the effect of snowpack disappearance on meltwater production, where production doubles as albedo falls, due to the transition from snow to ice. Although the foregoing adopt the surface energy balance approach outlined in eq. (4), some researchers report success by using a temperature index approach which incorporates potential radiation on slopes (Hock 1999).

Surface melt ecology

Life is associated with surface water on glaciers, just as it is elsewhere. Takeuchi and others (2001) have identified cryoconite holes on a Himalayan glacier as habitat for blue-green algae. Because of their dark colour, they suggest that this can substantially reduce the albedo of the glacier surface, a suggestion which Muller and others (2001) support as a consequence of their work on snow algal fields in Spitsbergen. Another observation from Spitsbergen is that just as algal cells can reduce albedo, thus enhancing melting, the meltwater streams which are produced will concentrate the cells. Thus the melt surface provides a habitat for life which itself conditions the ability of the surface to absorb the energy required for meltwater production.

ROUTING MELTWATER TO RUNOFF

Conceptually, one may consider total glacier surface meltwater generation, Q_o, to be the area integral of eq. (3) which, in finite difference notation is stated to be

$$Q_o = \Sigma_{j=1,n} M_j \qquad (4)$$

where n signifies the total number of melting surface elements which contribute meltwater to outflow. As stated, eq. (4) disguises the fact that there is an extremely complex set of pathways which connect surface meltwater production to outflow at the glacier terminus, R. The investigation and modelling of these pathways constitute what is currently the most exciting and intellectually stimulating area of glacier hydrology research.

One of the leading research sites for this research area is Trapridge Glacier, Yukon Territory, where Clarke (1996) has worked out a diagnostic modelling scheme to aid our thinking on how to visualize glacial flow pathways. The scheme is expressed by analogy to electric circuits, in which flows can be switched from one resistor to another. A crucial feature of the scheme is to view the resistors in terms of the relatively inefficient distributed flow paths of winter and the much more efficient conduit flow paths of summer, which are switched in as the englacial and subglacial flow network evolves during the melt season. Although the scheme was assembled specifically with Trapridge in mind, it could well apply to other glaciers because a key concept here is that the first switching of Q_o to summer flow does not initially allow diurnal variations to show up in R measurements. The diurnal signature in R appears after a second switching, from a poorly connected winter network, to the more efficient conduit flow of summer.

Englacial hydrology

As Fountain and Walder (1998) have noted in their review, englacial storage and movement of water is a key part of the flow system, firn storage alone exerting an important lag. They note that temperate alpine glaciers may store 200 mm of water, averaged over the glacier bed, with daily fluctuations of 20 to 30 mm, most of which may occur englacially. The importance of englacial flow is stressed by Wagnon and others (1998), in their study of Zongo Glacier, Bolivia, where both the englacial and the subglacial flow regimes are identified in outflow records. The importance of englacial flow is best identified during the early part of the melt season, before reorganization of the subglacial drainage system occurs (Gordon and others 1998).

The use of ground penetrating radar has recently provided information on englacial water storage. Moore and others (1999), in their examination of the hydrothermal structure of Hansbreen, Spitsbergen, find 1 to 2 percent water content for temperate ice, rising to 4 percent in wetter ice associated with crevasses and moulins. Murray and others (2000) find variations of water content with depth in Falljokull Glacier, Iceland: 0.23 to 0.34 percent at the glacier surface, 3 to 4.1 percent at 28 m depth, which corresponds to the piezometric surface, lower amounts at greater depths, falling to 0.09 to 0.14 at 102 m depth near the glacier bed.

At some point it become difficult to speak in terms of separate englacial and subglacial flow regimes. Oldenberger and others (2002) are clearly concerned about basal water pressure variations in their analysis of the borehole-acquifer system at Trapridge Glacier. This does raise questions, however, about how water pressure fluctuations control the movement of water in and out of englacial storage.

The subglacial system

Progress has been made in mapping the subglacial water drainage system through the use of radio-echo sounding on John Evans Glacier, Ellesmere Island, (Copland and Sharp 2001) and ground penetrating radar on Trapridge Glacier (Flowers and Clarke 1999). In the case of John Evans, warm/cold ice boundaries at the glacier bed have been identified. For Trapridge, the findings suggest that, at low water pressures, a dendritic drainage network exists that evolves into a more efficient morphology as water pressure rises.

The idea of a seasonal evolution in drainage morphology is represented in Cutler's (1998) modelling of subglacial tunnels for Storglaciaren, Sweden. Evolution in drainage morphology is also associated with glacier motion (Hanson and others 1998, Harper and others 2002). In fact, Kavanaugh and Clarke (2001) find three periods of strong basal motion at Trapridge Glacier, following establishment of a well connected subglacial drainage system. Basal water pressure is also important to understanding glacier flow, as Fischer and others (1999) suggest for Trapridge Glacier, associating the presence of so-called 'sticky spots' at the glacier bed with basal drag and slow motion. Flow speeds up at higher water pressures, as 'sticky spots' disappear and a lubricating film of water is established.

As Mair and others (2001) state, local variations in glacier motion cannot be explained simply by local variations in subglacial hydrology. Kavanaugh and Clarke (2000) find extreme pressure pulses at the base of Trapridge Glacier which cannot be explained by any known forcings. Mair and others (2002), working on Haut Glacier d'Arolla , suggest that the spring speed up of flow may be associated with the enlargement of subglacial cavities and the survival of residual cavities. Also, changes to the strength of the sedimentary layer at the glacier bed must be considered (Fischer and others 2001). Blake and Clarke (1999) show that the effects of this can be measured electrically, through changes in resistivity.

Outflow and water quality

The idea that a mid-summer switching occurs, from distributed to conduit flow, is supported by outflow measurements. Hodgkins (2001) identifies this as the point where diurnal cycling begins to appear in the outflow hydrograph and notes that the transition seems to occur rapidly. However, Anderson and others (1999) suggest that such flow transitions may occur several times during a melt season, as the subglacial conduit system extends up glacier. Stott and Grove (2001) also document minor fluctuations, of uncertain origin, which possibly may be associated with temporary damming of conduits as the local ice structure collapses.

As meltwater yield increases over the summer, so too does the sediment transport (Hodson and others (1998a). More revealing are the results of isotope analysis. For example, Tranter and others (2002) identify three types of subglacial water on the basis of isotopes from the outflow of Haut Glacier d'Arolla: solute rich from hydrologically inefficient flow, somewhat less solute from hydrologically more efficient flow and most dilute, but turbid, usually associated with borehole artefacts.

One of the key objectives in isotope analysis is to separate englacial water from subglacial, with a view to seeing how much of the surface melt is routed to outflow and how quickly. The evidence suggests that, once an efficient flow system is established, surface meltwater is dominant in the outflow records (Cecil and others 1998, Hodgkins and others 1998). As Mitchell and others (2001) point out, however, there are many minor, major and trace elements between the supraglacial and proglacial environments which can make the true picture difficult to see.

One of the keys to identifying outflow sources is to note that surficially routed snowmelt tends to be rich in atmospheric ions, such as Na^+ and Cl^-, while crustally derived solutes, such as SO_4^{2-} and Mg^{2+}, will appear in subglacial water (Wadham and others 2000). Certainly, the influence of rock is high at low discharges, but weakens as the discharge increases (Wadhan and others 1998). Snowpack oxygen isotope distributions, which may show marked vertical variation in the pre-melt season, tend to become homogenized as meltwater percolates through the pack during the melt season (Raben and Theakstone 1998).

ASSESSMENT AND OUTLOOK

In my previous report (Munro 2000), the focus had been on three main areas: mass balance change to reflect the state of the resource, winter accumulation to describe gains in water storage and the energy exchange associated with the hydrology of water yield from glacierized basins. It was noted that winter snow cover research from glaciers and ice sheets was relatively rare, a situation which now appears to be changing, as work from Greenland and Antarctica finds its way into the literature. Work on energy and surface meltwater generation is now such that separate sections on surface meltwater generation and routing to glacier runoff are warranted.

Canadian glaciologists continue to make important contributions to glacier hydrology, both in Canada and abroad. There is still the need stated previously (Munro 2000) to improve and expand the glacier mass balance network, an effort which is continuing under the direction of the Geological Survey of Canada (GSC) through its National

Glaciology Program (Koerner others 1999). A key advantage of GSC involvement is that it provides the opportunity for university partnerships in research, such that comprehensive research basins can be established. A key element of such basins is the ability to run mass balance programs with corresponding automatic weather stations and outflow gauging stations, such that year-round detailed records become available to study glacier hydrology throughout the year.

This raises the question of whether research efforts, which are scattered throughout Canada and abroad, should be concentrated in specific research basins. Such an idea has much to recommend it in terms of efficient use of research resources, though it risks some loss of creative independence on the part of researchers. Some sites, like Trapridge Glacier, must be studied simply because they are interesting. A more likely outcome is that the ideas which have been developed at the variety of research sites noted in this report will be incorporated into the field programs of key GSC glaciers. This will better inform the training of new glacier hydrologists at GSC sites, where it is important to interpret what changes in the annual mass balance of glaciers indicate about global change and how this will advance our understanding of a changing global hydrology.

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Recent (1999-2002) Canadian research on contemporary processes of river erosion and sedimentation, and river mechanics

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Abstract
This review is part of the Canadian quadrennial report to the International Association of Hydrological Sciences (IAHS), and focuses on the science and management aspects of sediment dynamics in rivers and drainage basins published between 1999 and 2002. The themes of this review were selected to be of interest to the hydrological sciences in general, and were chosen to represent a broad overview of the nature and directions of Canadian research in fluvial geomorphology, both in academia and in government and management. There is a large body of Canadian research that is aimed at elucidating the historical process record, for example through the use of lake sediments that reflect the erosional history of the contributing basin. This review, however, primarily concerns contemporary processes. The major themes of this review paper include sediment budgets and sediment yield; cohesive sediment transport; turbulent flow structure, sediment transport and bedforms; and bedload transport and channel morphology. These themes were selected because they have been the focus of substantial research in Canada.

Fluvial systems in Canada have a number of specific characteristics. First of all, Canada is a high latitude country, which means that, in most basins, the spring snowmelt is a dominant feature of the discharge and sediment transport regimes. Furthermore, in the northern part of the country, the presence of permafrost directly affects hydrological processes and is an important part of understanding fluvial processes and landforms. The northern location of Canada is also important from a historical perspective, since much of Canada was covered by ice during the Quaternary glaciations. As a result, the landscape in most of Canada is relatively young, and rivers are still actively adjusting to deglaciation, which only occurred during the late Pleistocene. The high latitude of the country also places it in that part of the world where CGMs generally indicate that the impact of global warming will be greatest in terms of temperature increase. Even though Canada is generally viewed as a relatively pristine country, its rivers are rarely unaffected by human activity. A multitude of dams has resulted in modifications of the annual discharge regime, and has led to changes in sediment storage and channel characteristics to a degree that is unknown, but likely substantial. Furthermore, in parts of Canada, human activity has led to a significant degradation of water and sediment quality—typically associated with urban, industrial and, sometimes, agricultural areas—and mobilization of large quantities of sediment within the drainage basin—typically caused by forestry and, sometimes, agriculture. Sediment quality and quantity in a stream directly affect the fish populations and, consequently, studies of the effect of human activity on fish habitat and behaviour form an important, practical part of fluvial geomorphology research in Canada.

Some of the characteristics of Canadian fluvial systems are reflected in the directions of research. Canadian research in fluvial geomorphology during the period of this review (1999-2002) continues in the same direction as earlier work summarized by Ashmore *et al.* (2000). There is a penchant for, and as a result, substantial progress has been made in, investigating the details of fluvial processes at relatively small scales. Examples of this emphasis are the investigations of floc structure, turbulence characteristics and bedload transport, which continue to form central themes in fluvial research in Canada. Translating the knowledge of small-scale, process-related research to an understanding of the behaviour of large-scale fluvial systems, however, continues to be a formidable challenge. Models play a prominent role in elucidating the link between small-scale processes and large-scale fluvial geomorphology, as they do in other fields such as climatology and oceanography. Canadian fluvial geomorphologists have recognized this role of models and, as a result, a number of papers describing models and modelling results have been published during the review period. It is to be expected that, in the future, the combination of detailed process measurements and models will gain importance in fluvial geomorphology in Canada, which will lead to an increased understanding of large-scale fluvial systems and strengthen the links between fundamental and applied research.

INTRODUCTION

Scope
This review is part of the Canadian quadrennial report to the International Association of Hydrological Sciences

(IAHS). Its focus is the science and management aspects of sediment dynamics in rivers and drainage basins published between 1999 and 2002. This review follows an earlier review of research published between 1995 and 1998 by Ashmore *et al.* (2000). The themes of this review were selected to be of interest to the hydrological sciences in general, and relate to the research area of the International Commission on Continental Erosion (ICCE), which is one of the scientific commissions of IAHS. The themes were also chosen to represent a broad overview of the nature and directions of Canadian research in fluvial geomorphology, both in academia and in government and management. There is a large body of Canadian research that is aimed at elucidating the historical process record, for example through the use of lake sediments that reflect the erosional history of the contributing basin. This review, however, primarily concerns contemporary processes. The major themes of this review paper include sediment budgets and sediment yield; cohesive sediment transport; turbulent flow structure, sediment transport and bedforms; and bedload transport and channel morphology. These themes were selected because they have been the focus of substantial research in Canada. The review ends with a section on development of techniques for research and monitoring, which presents recent Canadian contributions to methods and techniques.

SEDIMENT DYNAMICS AND SEDIMENT YIELD

Canada is the home of several large rivers, such as the Mackenzie, the Fraser and the St. Lawrence River, that drain substantial parts of the country and of the continent. These rivers have been the subjects of extensive investigations into their sediment transport characteristics. Rondeau *et al.* (2000) investigated the budget and sources of suspended sediment for the section of the St. Lawrence River between Cornwall, Ontario, near Lake Ontario, where the river starts, to Quebec City, Quebec, for the period 1989-1993. The results of this study indicate that Lake Ontario contributes less than 3% of the particulate load at Quebec City, whereas the tributaries on the south and north shores contribute 19 and 13%, respectively, of the sediment load. The remainder, nearly 65%, results from erosion of the bed and banks of the St. Lawrence River. The presence of fluvial or riverine lakes is a characteristic feature of the St. Lawrence River. These lakes—from upstream to downstream, Lake St. Francis, St. Louis and St. Peter—represent wide and shallow sections of the river rather than true lakes, and play an important role in the transport and storage of sediment and associated contaminants. Carignan and Lorrain (2000) used ^{210}Pb, ^{137}Cs and ^{7}Be to investigate the sediment dynamics in the fluvial lakes of the St. Lawrence River, with the objective to evaluate accumulation rates and characterize the mixed sediment layer. It was found that retention in the lakes ranged from 1.5% (Lake St. Peter) to 17% (Lake St. Francis) of the annual suspended sediment load. The relatively high mixing coefficient in the superficial sediments (14.9 ± 2.8 cm^2 year^{-1}), and the similarity between the annual particulate matter loading to the river and the mixed sediment inventory suggest that these lakes have a short memory of past conditions and can be expected to recover rapidly (within 2 to 5 years) following a decrease in contaminant influx. Lepage *et al.* (2000) investigated the sediment dynamics of the upstream portion of Lake St. Francis, and concluded that the suspended load on the northern side of the lake is mainly a function of the suspended load carried by the St. Lawrence River. In contrast, on the southern side of the lake, sediment resuspension and contributions of the local tributaries constitute an important portion of the suspended load. Lepage *et al.* (2000) calculate that wave action is likely to resuspend surficial sediments where the lake is shallower than 2 m, which accounts for a surface area estimated to be 32 to 35 km^2 between Cornwall Island and Thompson Basin. The St. Lawrence River system has been extensively modified for hydropower generation and navigation. Morin *et al.* (2000) applied bi-dimensional hydrodynamics to simulate past flow conditions and to produce quantitative descriptors of changes that have occurred in Lake St. Francis. A comparison of the pristine state, based on 1900 and 1870 measurements, with the present day geometry indicates significant changes in the morphology of the lake. Hydrodynamic simulations for the pristine and present day states indicate an increase of velocities over shoals and a decrease of velocities in deeper water for discharges less than 8800 m^3 s^{-1}. Dredging and straightening around Cornwall Island resulted in changes in the proportions of the discharge carried by the two channels around the island, with an increase in the flow through the south channel from 64 to 71% of the total river flow whereas flow in the north channel decreased from 36 to 29%.

Much of the sediment-related research on the St. Lawrence River has been carried out because of concerns about contaminant transport. Lean (2000) presents an overview of these issues. Filion and Morin (2000) investigated the effect of local sources on metal concentrations in littoral sediments and aquatic macro-invertebrates of the St. Lawrence River, near Cornwall, Ontario, and found that metal concentrations increased with the percentages of fines and organic matter, but were generally below the lowest effect level of the Ontario provincial sediment quality guidelines. Filion and Morin (2000) concluded from the spatial pattern of the metal concentrations in littoral sediments and that local sources of Hg and Zn had contributed to the contamination. Regarding bio-availability, Cr, Fe, Ni and Zn concentrations in macro-invertebrates were similar to or exceeded concentrations reported for deeper sites in the Cornwall area, despite the much lower concentrations in littoral sediments, suggesting that bio-availability of these metals is greater in littoral than in deeper sediments. Despite the effect of local point sources, a comparison with the Ontario sediment quality guidelines and with other sites in the Great Lakes - St. Lawrence system suggests that metal contamination of littoral sediments and invertebrates along the investigated section of the river was relatively low. The fate of sediment in the St. Lawrence River is not just affected by the flow in the river itself, but also by the flow conditions in the tributaries. Lepage *et al.* (2000) suggest that fluctuations in the winter discharge of the south shore tributaries contribute to sediment resuspension and redistribution of contaminants such as mercury and polychlorinated biphenyls in Lake St. Francis.

On the other side of the continent, Sichingabula (1999) investigated the magnitude-frequency characteristics of effective discharge for suspended sediment transport in the Fraser River, British Columbia, using discharge and sediment concentration data for the period of 1965 to 1988 that were collected and archived by the Water Survey of Canada. Sichingabula (1999) defines the effective discharge as the mid-point of the discharge class transporting the greatest portion of the suspended sediment load. The analysis indicates that the effective discharge concept is applicable to the Fraser River basin. Sichingabula (1999) presents equations for predicting the class-based effective discharge in the Fraser River basin from bankfull discharge and drainage area. McLean *et al.* (1999a, 1999b) investigated sediment transport along the lower Fraser River using sediment transport data collected between 1966 and 1986 in combination with channel surveys conducted 32 years apart. The results of this study indicate that within-reach variations in transport are substantial, so that the results from a single cross-section may not be representative for the reach. Furthermore, the authors conclude that, most of the time, the river's sediment load is far less than its hydraulic capacity, and that approaches based on morphological change seem to be most cost-effective and can best take advantage of available historical information.

Church *et al.* (1999) investigated fluvial clastic sediment yield in Canada, and present maps of regional sediment yield for standard areas of 1, 10^2 and 10^4 km^2. The specific sediment yield increases downstream in most regions, indicating regional degradation of river valleys. Aggradation on a regional basis, however, is taking place in the southern prairies, whereas specific sediment yields on average are similar at all scales in southern Ontario. Evans and Church (2000) address an issue that is crucial for studies involving lake sediment-based reconstructions of sediment yield, i.e., how to derive error estimates of the estimated sediment yield. Evans and Church's (2002) approach involves modelling the physically controlled, spatial variability of sedimentation using regression surfaces fitted to point values of sediment mass derived from multiple cores. Deviations from these surfaces are interpreted to represent the remaining, unstructured variance, which provides an error estimate.

COHESIVE SEDIMENT TRANSPORT

The transport of cohesive sediment in aquatic systems is characterized by interactions among fine-grained sediment particles that cause flocs to form. Flocs have relatively low densities, large pore spaces and reactive surfaces that remove contaminants from the water column. Flocculation is an important mechanism for particle removal in streams, lakes and oceans that alters the hydrodynamic characteristics of solids by changing the density, porosity, settling velocity and surface area. Numerical models designed to simulate contaminant transport, fate and bioaccumulation in aquatic environments have begun to include a cohesive sediment transport component but a better understanding of cohesive sediment transport processes (erosion, deposition, flocculation) is required to improve model predictions. The following sections review recent studies that have advanced knowledge regarding the nature and transport of cohesive sediment in aquatic systems.

Nature of Cohesive Sediment

Cohesive materials represent variable proportions of the total annual sediment flux in many Canadian rivers (Stone and Saunderson, 1992; Droppo *et al.*, 1998; Krishnappan, 2001) and urban storm water runoff (Droppo *et al.*, (2002). The morphology and settling characteristics of these materials vary in response to physical, chemical and biological attributes of individual rivers and sediment sources (Petticrew and Droppo, 2000). Cohesive suspended sediment is commonly transported in fluvial systems in a flocculated form, and many larger flocs do not settle within the Stokes' region of Reynolds numbers (Droppo *et al.*, 2000). In a study of river and lake sediment, Droppo *et al.* (1999) reported that only flocs < 100 μm (equivalent spherical diameter) settled within the Stokes' region (Re < 0.2). The densities of these flocculated materials ranged from 1 to 1.4 g cm^{-3} but the majority of flocs had densities of less than 1.1 g cm^{-3}. Floc porosity increases with floc size, and low floc densities are caused by the entrapment of water in the pore spaces of flocs (Droppo *et al.*, 2000).

Flocs consist of a complex matrix of microbial communities, organic particles, inorganic particles, inter-floc pore spaces, and interstitial water (Droppo, 2001). Advances in understanding the structural components of a floc and its individual properties have led to the development of a conceptual model that links both the structural and behavioural components of flocs (Droppo, 2001). The model redefines the traditional view of suspended solids as discrete particles, to a collection of compositionally diverse flocculated particles that behave as individual micro-ecosystems with complex physical, chemical and biological behaviours.

Fractal dimensions have been used to quantify the morphology of particle populations formed in different fluid mechanical (Jiang and Logan, 1991; Logan and Kilps, 1995), stream (De Boer, 1997) and marine environments (Logan and Wilkinson, 1990). Fractal dimensions reflect the nature of particles and their mechanism of formation and particle properties such as settling velocity and density are a function of their fractal dimensions. De Boer and Stone (1999) examined the fractal dimensions of suspended solids in streams to compare settling and filtration sampling techniques for particle size analysis. Systematic differences between the two methods were observed but the filtration method was more sensitive to indicating differences within and between the sites in two basins. De Boer *et al.* (2000) investigated the fractal dimensions of individual flocs and floc populations of suspended solids collected during snowmelt in southern Ontario streams with contrasting riparian buffer zones. Fractal dimensions of both individual flocs and floc populations provided similar information about temporal changes in sediment source contributions and about the contrasting effectiveness of the riparian buffer zones in the two basins.

Stone and Krishnappan (in press, a) examined the fractal dimensions of particle populations of cohesive river sediment in a rotating circular flume and used image analysis to evaluate the structure and size distribution of flocs formed during the deposition process at four conditions of steady state flow. As shear stress increased from 0.058 to 0.121 Pa, particle boundaries became more convoluted and shape irregularity of larger particles increased compared to the smaller ones. Micro-flocs were the building blocks of the larger flocs suspended in the water column and the stability of larger flocs was a function of the shear stress at steady state. Stone and Krishnappan (in press, b) determined the fractal dimensions of particle populations of cohesive sediment during settling experiments in an annular flume with different initial conditions at a constant bed shear stress. The ratio of initial and steady state sediment concentration for both runs was 0.54 and is a function of bed shear and not the initial sediment concentration. Fractal dimensions (D, D_1, D_2) were not significantly different for the two experimental runs at steady state ($t = 300$ minutes).

An international symposium on The Role of Erosion and Sediment Transport in Nutrient and Contaminant Transfer was held at the University of Waterloo in 2000. Proceedings from the symposium advance knowledge of erosion and sediment transport processes in relation to chemical transfer at a range of spatial and temporal scales (Stone, 2000). Ongley and Droppo (2000) provide guidance on sampling suspended solids for water quality investigations.

Cohesive Sediment Transport

Numerical models have been developed to predict the transport and fate of sediment and contaminants but they require information on the transport characteristics of sediments as input parameters. For cohesive sediment, variables such as erosion rate and critical shear stress for erosion and deposition must be determined by direct measurement. The following section reviews recent laboratory and field investigations that have advanced knowledge of cohesive sediment transport.

Laboratory Studies: Flocculation of cohesive materials and settling of flocs on the river bed result in the formation of surficial fine grained laminae [SFGL] (Droppo and Stone, 1994) that represents a significant potential sink for contaminants bound to cohesive sediment (Stone and Droppo, 1994). Several recent studies have advanced knowledge regarding the formation and erosional characteristics of SFGL. Droppo *et al.* (2001) conducted experiments in an annular flume using commercially available kaolinite clay and contaminated bed sediment from Hamilton Harbour to assess the effect of depositional history on the stability of contaminated bed sediment. Results of the study demonstrate that bed strength (erodibility) is dependent on both the degree of bio-stabilization and the flow conditions under which the bed is deposited. In a related study, Lau and Droppo (2000) report that the critical shear stress for beds deposited under shear was up to eight times larger than for beds deposited under quiescent conditions. In a series of sequential erosion/deposition experiments, Lau *et al.* (2001) demonstrated the effects of depositional history on sediment erosion and showed how the rate of erosion and the amount of sediment eroded reflect the structure of the bed and flocs that formed it. Their research shows that layers of sediment deposited with different depositional history will not have the same shear strength and therefore similar flow conditions will not necessarily produce the same erosion rates.

Using a rotating circular flume, Krishnappan and Marsalek (2002a) measured the transport characteristics of cohesive sediment deposited in an on-stream storm water management pond. The critical shear stress for deposition (0.050 Nm^{-2}) and erosion (0.12 Nm^{-2}) of pond sediment were determined and used to develop empirical relationships to estimate sediment deposition and erosion as a function of shear stress. A new model to predict transport characteristics of sediment from an on-stream storm water management pond was developed by Krishnappan and Marsalek (2002b). Skafel and Krishnappan (1999) investigated the depositional characteristics of mud from Port Stanley harbour using a rotating annular flume and showed that duration of sample storage, presence of bacteria and textural composition of the sediment affected depositional behaviour.

Millburn and Krishnappan (2003) carried out an intensive field program before river-ice break up and conducted controlled experiments in a rotating annular flume to determine the critical shear stress for erosion and deposition of Hay River sediment. They proposed a modelling strategy for analyzing the under-ice transport of cohesive sediments in the Hay River. Krishnappan (2000) developed a new algorithm for the transport of fine sediments in the Athabasca River based on laboratory experiments in a rotating circular flume. Accounting for differences in the critical conditions for erosion and depositional processes of fine sediment, the algorithm was incorporated into and improved the performance of the contaminant transport model (WASP5). These algorithms have been used to improve accuracy of the models RIVFLOC and FINSED, initially developed by Krishnappan (1991, 1997).

Field Studies: A series of field studies have been conducted with portable flumes to determine the *in situ* transport properties of cohesive sediment in lakes and rivers. Droppo and Amos (2001) used an *in situ* annular flume to assess the effect of shear stress on the structure and stability of bottom sediments in Hamilton Harbour. They developed a general three-layer model that depicts organic flocs of the fine-grained surface layer (Layer 1), compressing within a collapse zone (Layer 2) to form a consolidated bed (Layer 3). The structure of eroded materials evolved from low-density flocs from the fine-grained surface layer to dense aggregates of the consolidated bed. In a related study, Amos *et al.* (2003) compared three methods to estimate the threshold shear stress (τ_c) of lakebed sediment using the benthic flume Sea Carousel. The method, which extrapolates a regression of suspended sediment concentration and fluid transmitted shear stress, is recommended for evaluation of the erosion threshold conditions.

Krishnappan (2000) used a submersible laser particle-size analyzer to show that suspended solids in the Fraser River downstream of a pulp mill outfall were transported as flocs and that fibrous organic material in the effluent promoted flocculation of inorganic solids suspended in the water column. Flocculation of suspended solids by pulp mill effluent increased the deposition rate of sediment in the river. In a year-long evaluation of the structure and composition of suspended sediments upstream and downstream of a point source of pulp mill effluent discharged into the Fraser River, Petticrew and Biickert (1998) and Biickert (1999) found no significant differences in the fractal dimensions and only a very small increase in the d_{84} of the floc populations downstream of the effluent. The significant seasonal variation in floc size far exceeded the near-field effects of the pulp mill effluent. As well, no significant differences between the upstream and downstream sites were noted in the total amounts size and deposition rate of fine sediment collected in gravel traps. While this result differs from the findings of Krishnappan (2000), the differences can be explained by the use of different sampling techniques and by the scale at which the effect of the effluent was evaluated. Krishnappan (2000) tracked the plume downstream using an in-situ laser particle analyser, whereas Biickert (1999) used field collection and laboratory microscopy and sampled at stationary sites approximately 300 and 600 m downstream of the effluent pipe.

Subsurface drainage represents a relatively unknown component of low-order stream sediment transport and little is known about the particle properties and fluvial transport characteristics of tile sediment in relation to varying soil texture, land use and moisture conditions. Stone and Krishnappan (2002) examined the effects of irrigation on tile sediment transport in a headwater stream. Sediment yield from the controlled irrigation event was 4.6 kg ha^{-1} and tile sediments were fine-grained ($d_{50} \approx 5$ μm). Flow in the study reach was modelled with MOBED to determine the bed shear stress and relate the size characteristics and degree of flocculation to previously conducted flume studies (Stone and Krishnappan, 1997).

Shantz *et al.* (2003) examined the effect of drawdown on the timing and magnitude of suspended solids and associated phosphorus export from a reservoir located in an urbanized southern Ontario watershed. Suspended solids and phosphorus export increased significantly with decreasing lake level and contributed to the annual particulate P flux to Lake Erie. Stone and Haight (2000) used a field-based water elutriation system to investigate the occurrence and distribution of dioxins and furans separated size fractions of suspended sediment during a spring storm event in Canagagigue Creek near Elmira, Ontario. Several furan and dioxin compounds including 2,3,7,8 – T_4CDD and 2,3,7,8 – T_4CDF were detected in suspended solids at comparable levels to those reported in previous investigations of creek bottom and floodplain sediments. There was no significant relationship between grain size and organic contaminants in suspended solids because of the highly flocculated and bio-stabilized nature of river bottom sediment eroded from the study site.

In a study of suspended solids, trace metals and PAH concentrations from coal pile runoff to Hamilton Harbour, Curran *et al.*, (2000) found that trace metal concentrations often exceeded the Canadian Water Quality Guidelines for the Protection of Aquatic Life and concentrations of some PAHs exceeded the provincial "Severe Effect Level." They recommended that coal pile discharge be treated as a remedial action for the harbour. Lévesque and De Boer (2000) investigated the trace element chemistry of surficial fine-grained laminae in the South Saskatchewan River. Concentrations of Cu, Zn, Cd, Pb and U in SFGL samples downstream of the city of Saskatoon were significantly greater than upstream samples collected on days when the flow velocities were low.

TURBULENT FLOW STRUCTURE AND SEDIMENT TRANSPORT

Context

Flow provides the impetus for sediment transport and the development of bedforms in river environments. The study of flow turbulence has become increasingly relevant over the last few years for two reasons. First, sediment transport has been found to be highly variable in space and time (Thorne *et al.*, 1989; Nelson *et al.* 1995). This reduction in scale means that the relevant flow parameters for its study also must reduce in scale. It is necessary to look beyond mean flow parameters. Second, continued investigations into the structure of turbulent flow have revealed that is organized on a variety of scales (Roy *et al.*, 1996). There thus exists a strong potential for feedback mechanisms between the organization of flow and the organization of the bed. Recent research in Canada has contributed to the study of the character of turbulent flow in rivers and of the interaction of turbulence with sediment transport and bedforms.

Flow Structure in Gravel-Bed Rivers

Buffin-Bélanger *et al.* (2000a) documented and visualized large-scale flow structures in a gravel- bed river. Velocities were measured using an array of Electromagnetic Current Meters sampling at a rate of 20 Hz and flow structure was visualized using a new technique based on time-space velocity matrices. Large-scale flow structures were shown to consist low and high-speed wedges and were a dominant feature that occupied the full depth of flow. The structures were clearly associated with large peaks in bed shear stress. Lawless and Robert (2001a) obtained high-resolution, three-dimensional data around pebble clusters in a flume to continue investigation into the effects of pebble clusters on turbulent structures. They found high lateral variance of flow properties but were unable to confirm the hypotheses of a standing horseshoe vortex. Instead, results suggested an intermittent structure that converged in the upwelling zone downstream of a cluster. In a second paper, Lawless and Robert (2001b) addressed the effects of pebble clusters on local and average velocity profiles. The increase in the scale of roughness induced pressure gradients that changed the shape of the velocity profiles and increased the average shear stress in the outer region by 100%. Zimmerman and Church (2001) examined the stability of step-pool structures in mountain streams. The stability of the structures was found to be primarily related to particle interactions and the formation of very stable structures transverse to the flow. In their discussion of flow hydraulics, however, an empirical relation between step height and the depth of the downstream pool was found. This suggested that flow turbulence was playing at least a modifying role as the dissipation of momentum in the pools caused sediment to be moved out and deposited downstream. Millar (1999) looked at the modification of energy dissipation due to form roughness in gravel-bed rivers. Compiling data from a number of sources, he demonstrated that overall roughness can deviate strongly from widely used relations to grain size due to roughness generated by bedforms. In an effort to integrate recent advances of the understanding of turbulence in gravel-bed rivers, Buffin-Bélanger *et al.* (2000b) developed a conceptual model of the interactions between bursts, as classically defined in boundary layer dynamics literature, surface boils, shedding motions from clusters and protruding particles, and large-scale flow structures. These schematics, while speculative, highlight the range of scales in coherent turbulent structures and their complex interactive potential. The role of turbulence in models of particle interactions has not been explicitly accounted for. Tribe and Church (1999) and Malmaeus and Hassan (2002) developed two-dimensional plan models that simulated individual particle interactions and looked for the development of surface structure. In the Tribe and Church model, particles only interacted when directly in contact with each other, ascribing no role to the local flow dynamics. The model of Malmaeus and Hassan introduced the concept of a resistance field around particles. As no direct contact was required for particles to interact, this concept implicitly represented the modification of turbulent flow properties around protruding clasts and particle clusters.

Flow Structure over Dunes

Robert and Uhlman (2001) undertook a comparative study of flow turbulence characteristics above three bed types. The beds utilized were fixed positive casts of fluvial forms in a sand-bedded flume and represented the transition of a bed from a ripple to a dune-dominated morphology. An Acoustic Doppler Velocimeter (ADV) was used to measure velocities in three dimensions at 10 Hz. A gradual increase in overall turbulence intensity was observed through the transition and the spatial variability of turbulence increased in a non-linear fashion, with dunes inducing very high Reynolds' stresses at some locations on the bed. Villard and Kostaschuk (1998) investigated the effect of dune geometry (i.e., symmetric versus asymmetric dunes) on the relation between shear velocity and suspended sediment over large dunes in the estuary of the Fraser River in British Columbia. They found that the roughness length for the asymmetric dunes was much larger than for the symmetric dunes. Asymmetric dunes were found to be relatively inactive remnants of symmetric forms developed during high flows. Kostaschuk (2000) continued the investigation of turbulence around large dunes in the Fraser River by combining velocity measurements from an ECM and optical backscatter (OBS) probes, both taken at 1 Hz. On the stoss side of the dunes, the near-bed velocity and sand concentration increased, while in the leeside separation zone the intensity of turbulence increased and reversed mean flow was observed. Wake-flapping and eddy shedding from the separation zone were found to be dominant sources of turbulence generation. Both processes advected eddy structures into the ambient flow at an angle of 23-25° from the horizontal. The importance of the dune structure for sediment transport was investigated by Villard and Church (2003). They repeatedly surveyed the bathymetry through the middle of a navigation channel in the Fraser River and found dune dimensions to increase in height and length in response to increases in flow, though lags between flow and dune dimension peaks were commonly observed. The study was able to confirm that dune-associated transport corresponds with overall bedload transport in the channel, highlighting the importance of understanding dune mechanics. Bedform mechanics in sand-bed environments are also active areas of research in other earth science fields. Recent progress in aeolian sand transport and dune formation is reviewed by Walker and Nickling (2002), while Crawford and Hay (2003) present the results of an investigation into ripple migration in near-shore ocean environments. These fields offer ample opportunity for sharing results, as the forms being studied are similar, though each environment presents unique challenges.

Flow Structure at River Confluences

Recent research into the flow dynamics at river confluences has emphasized both the role of turbulence and the advantages of numerical simulations. DeSerres *et al.* (1999) conducted a field investigation into the three-dimensional flow structure at the Bayonne-Berthier confluence in Quebec. Velocity measurements were obtained at 20 Hz in three-dimensions using Electromagnetic Current Meters (ECMs), though limitations of the technology meant that only two dimensions could be measured simultaneously. The study documented the structure of the flow at the confluence over a range of flow stages and bed conditions. Results illustrated the feedback from bed morphology to the flow structure and the dominant role of vortices in the mixing layer in controlling bed morphology and sediment transport. Lane *et al.* (1999) began a series of articles with a critique of the manner in which secondary flow circulation is typically defined at channel confluences. They argue that secondary flow can appear mathematically where in fact, due to flow convergence and the downstream transfer of water, no secondary circulation occurs. These arguments were developed more fully in Lane *et al.* (2000) and the dangers demonstrated using numerical simulations. The simulations resolved larger scale eddies using the Navier-Stokes equations, while smaller, sub-grid scale turbulence was resolved using the SIMPLE algorithm of Patankar and Spalding (1972). Results were verified by comparing with flume and field data. One of the main problems found was the requirement of data rotation for secondary flow identification. Results were shown to be radically different depending on the definition of rotation used. The research also showed the dangers of working with mean flow values, as instantaneous values indicated the dominance of periodic eddy shedding events rather then a closed secondary circulation cell. Further simulation work in Bradbrook *et al.* (2001) investigated the roles of bed discordance, junction angle, and velocity ratio at river confluences. Flow separation as a result of bed discordance was found to result in upwelling into the separation zone and significantly reduced mixing lengths compared to confluences without bed discordance. Biron *et al.* (2002) conducted a field-based study to obtain accurate measurements of variations in water surface in the Bayonne-Berthier confluence. They were able to document coherent patterns of super-elevation and depression including super-elevation of the mixing layer. This information was incorporated into numerical simulation models and found to increase the precision of simulations.

Sediment Transport and Flow Turbulence Interactions

The initiation of sediment transport by flow turbulence has been investigated by a number of authors in the last decade (e.g. Nelson *et al.*, 1995). What remains poorly understood, however, are feedback mechanisms such as the modification of the turbulence by entrained sediment. A few recent flume studies in Canada have aimed to address this problem. Li and Gust (2000) looked at drag reduction due to the introduction of cohesive sediments. They designed a series of flume experiments where kaolinite was fed into flows over a smooth bed at different concentrations. Skin friction shear velocities and velocity profiles were measured using an array of hot-film sensors. They found that shear velocity was reduced up to 70% as a result of the cohesive sediment. The decrease in shear velocity was attributed to a thickening of the inner wall layer and turbulence damping. Bergeron and Carbonneau (1999) investigated the effect of sediment load on flow properties by feeding sediment into a flume with a fixed rough bed and maintaining flow conditions above the entrainment threshold. Velocity profiles were measured and shear velocity, roughness length, and resistance to flow were calculated. It was found that mean flow velocity decreased, while shear velocity, roughness length and resistance to flow increased. A shear velocity plateau was reached which corresponded with the sediment transporting capacity of the flow. In a continuation of the work, Carbonneau and Bergeron (2000) found that under some conditions, the introduction of sediment resulted in faster mean velocities. These conditions corresponded with situations in which flow turbulence decreased. The authors argue that the effect of sediment load on roughness is difficult to predict because parameters such as roughness are representative of energy dissipation on a large scale. The incorporation of greater spatial and temporal variability through the study of turbulent kinetic energy and turbulent dissipation is necessary to identify the active physical mechanisms.

Fish Habitat

A gravel-bed river presents a heterogeneous habitat for fish. Local scale variation in turbulence and sediment transport will have strong implications for fish energetics and success of fish reproduction. Two recent studies have looked at the spatial variability of scour and impacts on fish spawning redds. Lapointe *et al.* (2000) delineated potential spawning areas in a river and then completed high-resolution topographic surveys to determine impacts of three significant floods on scour and fill in the delineated zones. The data suggested that the probability of scour ranges from under 5% for a typical spring flood to approximately 20% for a rare event with a recurrence period measured in centuries. Rennie and Millar (2000) used a spatially dense array of scour chains in a short spawning reach to determine whether scour depths in spawning redds are different from the surrounding bed. They found the spatial variability of scour to be greater than the density of their measurements and no statistically significant difference between the redds and the adjacent bed. Within the four egg pockets instrumented, however, none were scoured to the critical depth during a typical spring flood, while scour was much greater in the redd tailspill areas.

BEDLOAD TRANSPORT AND CHANNEL MORPHOLOGY

Channel morphology is governed by the volume and distribution of water flow, volume and texture of sediment supply to channels, texture of bed and banks material and riparian vegetation. Recent Canadian research has addressed four major issues; bed surface structures that regulate sediment transport, morphological methods for the estimation of sediment transport, channel morphology and response to changes in the governing conditions, and modeling of channel dynamics. The current and future research will likely be focused on the role of surface structures in stabilizing channels, flow characteristics and sediment transport, channel morphology of small forested streams, modeling of sediment transport and channel morphology, and watershed scale sediment storage and transfer.

Sediment transport and bed surface structures

Sediment transport fundamentally consists of movement of individual particles. The motion of grains is not continuous, but rather consists of a series of steps and rest periods. The apparent bedload velocity was measured by Rennie *et al.* (2002) with acoustic Doppler current profiler bottom tracking. The mean bedload velocity measured at sampling stations correlated well with mean bedload transport rates in the gravel-bed reaches of the Fraser River (Rennie *et al.* 2002; Rennie and Miller, in press)) as well as the sand-bed reaches (Rennie and Villard 2003).

Bed material transport in gravel bed rivers commonly occurs at near-threshold transport rates. Under these conditions, the bed is only partially mobilized and most of the bed material remains immobile for extended periods of time (e.g., Wilcock and McArdell, 1993). Hassan and Church (2001) and Church and Hassan (2002) conducted a detailed bedload study in Harris Creek. Fraction transport rates in Harris Creek plot up to three orders of magnitude below the reference transport rate corresponding to the threshold of movement suggested in the literature (Church and Hassan, 2002). Sediment mobility is largely controlled by the bed surface structures; they stabilize the bed and reduce the transport rate by orders of magnitude (Church *et al.*, 1998; Church and Hassan, 2002). The sensitivity of bedload flux to variations in flow strength was examined via trap-specific rating curves (Hassan and Church, 2001). All the rating curves are very sensitive, indicating that bedload flux remains in the regime of partial transport. The ratings also exhibited seasonal hysteresis and vary from trap to trap.

Lawless and Robert (2001a) conducted flume experiments to investigate the effects of pebble clusters on flow characteristics. Their results demonstrate the relative importance of the lateral flow component in the development of turbulent flow structure. In a second paper, Lawless and Robert (2001b) addresses the effects of increasing scale roughness on local and average velocity profiles. Imposing pebble clusters on plane bed increased the average shear stress in the outer flow region by 100% from the plane bed conditions. Bergeron and Carbonneau (1999) and Carbonneau and Bergeron (2000) conducted flume experiments to examine the effect of sediment concentration on bedload roughness and turbulent properties of the flow. At high sediment concentration the whole velocity profile become affected by bedload and that the importance of this effect always remains larger near the bed. Church *et al.*, (1998) and Hassan and Church (2000) conducted flume experiments to examine the development of surface structures and explore their influence on channel stability. The zero feed experiments show that such structures developed simultaneously with the armour (Church *et al.*, 1998). Under varying feeding conditions, the bed surface structures remained intact during the experiments, but the bed surface composition fined with increasing sediment transport rate (Hassan and Church, 2000). Working in small steep stream, Zimmerman and Church (2001) identified three distinct populations of sediment, each exhibiting a characteristic degree of stability; the immobile boulders that form the step keystones, loose cobble population in the pools, and fine pocket deposits which is the most mobile under the current flow regime.

Scour and fill

Annual patterns of scour and fill in the Squamish River were examined by Paige and Hickin (2002) using sonar cross-sectional surveys. The analysis shows no simple relation between mean bed elevation and discharge. The bed-elevation regime is dominated by the variation in sediment transport rate, which in turn depends on the sediment supply. The spatial variation of bed scour and fill depths was observed for six flood periods in a short gravel-bed spawning reach of Kanaka Creek, British Columbia (Rennie and Millar 2000). Despite an intensive array (0.059 monitors per m2) of scour monitors, scour depths were not spatially autocorrelated, suggesting that monitoring was not sufficiently intensive to capture spatial variability of bed movement. Fassnacht and Conly (2000) used bathymetric evidence to examine the spatial and temporal stability of scour hole on the East Channel of the Mackenzie Delta. The study shows that the hole remained stable even during the record event of 1988, however some lateral erosion and sedimentation has been recorded.

Morphological methods for sediment transfer

Recently, there has been growing emphasis upon using changes in the channel morphology to estimate sediment transfer in rivers (Ashmore and Church, 1998; McLean and Church, 1999; McLean *et al.*, 1999). Using a small-scale model, Lindsay and Ashmore (2002) examined the magnitude of bias associated with estimates of scour and fill that are derived by differencing topographic surfaces. Their analysis showed that the relation between measured cumulative volumes of scour and fill and survey interval could be described using inverse functions. Ham and Schwab (1998) examined erosional and depositional trends along a 1990 km2 wandering gravel-bed channel of the Kitimat River, British Columbia. In a second paper, Ham and Church (2000) examined the relation between planimetric channel changes and bed-material transport on a 1230 km2 wandering gravel-bed channel of the Chilliwack River, British Columbia. Morphologic change and sediment transport in the system is dominated by the passage of large floods, which cause extensive bank erosion and introduce much of the supply of available bed material to the channel. Similarly, Eaton and Lapointe (2001) used the geomorphic method and Meyer-Peter Muller formula to estimate sediment transport during two extreme floods on the cobble-bed Sainte Marguerite River, Quebec. They attributed the channel stability to changes in channel pattern and the initiation of bed degradation following channel rectification during the 1960s. Gomi *et al.* (2001) assessed that distribution and accumulation of woody debris and sediment related to management and disturbance regimes in headwater channels. They found that number of woody debris pieces was significantly correlated with the volume of sediment stored in channels.

Morphological thresholds, Sediment transfer, and storage

The morphological transitions and thresholds associated with sediment and water transfer through riverine landscapes were reviewed by Church (2002a). Church discusses the influence of stream competence on the distribution of sediment and channel characteristics through the drainage basin. In a second paper Church (2002b) addressed the question of sediment transfer in cold regions. He suggested that sediment transfer through a drainage basin on long time scales (102-106) is mainly matter of sediment storage. Church argued that sediment in a drainage basin moves through a cascade of linear reservoirs. Church points out that the glaciations of the Pleistocene Epoch as the most significant sedimentary disturbance event in geologically recent times in Canada. The legacy of the glaciations is still present (Church and Slaymaker, 1989; Church *et al.*, 1999) and detectable in landscapes that have been recently been extensively disturbed by human activities (Church, 2002b). The transport-storage relations of sediment reservoirs were also investigated by Lisle and Church (2002). Their analyses lead to the proposition of a two-phase conceptual model: phase I in which filled channel response to variation in sediment supply by changes in the volume of stored sediment and in phase II channel mobility is response to sediment supply through armouring and roughness.

Channel morphology and adjustment

Channel steps, most commonly formed by boulders and woody debris, are significant geomorphic units, affecting the stability and sediment transport capacity in steep headwater channels. Channel morphology in headwater systems has been characterized by Zimmerman and Church (2001). They show that the step-pool sequence undergoes a considerable change in hydraulic characteristics between low and high flows. Gomi (2001) and Gomi *et al.* (2003) found that the timing of clear cutting and mass movement modified recruitment of woody debris pieces: this altered the distribution of channel steps and reach morphology along headwater channels. Then, based on detail monitoring and observation of sediment transport, Halwas and Church (2002) presented a classification and description of channel units in seventeen small, high gradient mountain channels. Channel morphologies were described according to their bed slope, and dominant bed material texture and organization.

Characteristics of bedforms in sand bed rivers were the focus of studies in Canada. Prent and Hickin (2001) described statistical character of bedforms and examined the relation among discharge, flow resistance, and channel morphology of the Lillooet River in British Columbia. Dune characteristics in the estuarine Main Channel of the Fraser River were investigated by Villard and Church (2003). Then they used migration rates estimates and geometry to estimate sediment transport. Ekes and Hickin (2001) explored the use of ground penetrating radar (GPR) to study sedimentary architecture of alluvial fans. This is the first application of the GPR to the study of the complex internal structures of alluvial fans.

Historical downstream in channel geometry, grain size, and gradient of the Sunwapta River, a proglacial stream in Alberta, were examined by Chew and Ashmore (2001). The analysis was used to test empirical hydraulic geometry relations and rational regime equation predictions of channel adjustment. The multiple response of the Saint-Marguerite River to large scale meandering rectifications was investigated by Talbot and Lapointe (2002a). In terms of reestablishing sediment transport equilibrium along the river, the two most important responses were vertical reprofiling and pavement (coarsening) of bed surface sediment. The influence of bank vegetation on alluvial channel pattern was investigated by Millar (2000). He developed a planform stability diagram to determine the sensitivity of gravel-bed rivers to changes in bank vegetation.

In July 1996, a severe rainstorm caused widespread, catastrophic flooding along many rivers in southern Quebec, but especially along the tributaries of the Saguenay River in the Chicoutimi and Lac-Saint-Jean region. Brooks and Lawrence (2001) documented the geomorphic effects of these floods. Along some reaches, channel widening and floodplain erosion resulted in transforming the river from meandering to braided. The most significant geomorphic effects along some of the study reaches occurred at run-of-the-river dams. Four of these dams were overtopped by the floodwaters, resulting in rapid and deep erosion into the unconsolidated sediments next to the dams and the formation of new channels that captured the flow of the river. Floodwaters also overtopped a fifth dam within an urban subdivision, causing scour of the overburden and roadbeds, and damaging and destroying buildings. Gullies and scour holes eroded into flood plain meandering and braided channels in northern rivers were investigated by Smith and Pearce (2002). The two anomalous fluvial landforms are caused by ice jams. Makaske *et al.* (2002) investigated floodplain sedimentation rate, channel avulsion and evolution of the anatomizing reach of the upper Columbia River. The study confirms Makaske's (2001) characterization of the river as internally dynamic anastomosing system. Finally, the impact of climatic changes on rivers and river processes in Canada has been discussed extensively by Ashmore and Church (2001).

Models

Martin and Church (2000) presented several-revised version of Bagnold's original formula. In order to obtain consistent empirical representation, the revised versions were calibrated using a wide range of data. On the basis of dimensional analysis, Martin and Church (2000) developed a rational form of the Bagnold equation.

Tribe and Church (1999) developed a simulation model based on mutual interactions between particles on a streambed. Simulating individual grain displacement by designing a number of rules and statistical functions for entrainment, movement and entrapment, they succeeded in reproducing patterns observed in Harris Creek, British Columbia. A similar approach was taken by Malmaeus and Hassan (2002); they developed two-dimensional stochastic model that simulates the movement of individual particles on a bed that exhibits small-scale surface structures and has a riffle-pool

Several one-dimensional models of fractional sediment transport and deposition have been developed in the last few years (Ferguson *et al.* 2001). Ferguson *et al.* (2001) applied the SEDROUT to study the development of channel longitudinal profile and sediment fining along the Vedder River, British Columbia. Talbot and Lapointe (2002b) used the same model to simulate the response of the Sainte-Marguerite river over the last 32 years to meander straightening. The study shows that pavement coarsening after rectification buffers the system against extreme degradation which appears to limit the extent of propagation of degradation upstream of the straightened reach (Talbot and Lapointe 2002b). Two-dimensional computational model for the simulation of transient variation in bed topography was developed by Fares (2000). The model combines the longitudinal flow momentum with the continuity principle of the sediment transport. The simulated bed level profiles were compared with field data. Rice and Church (2001) tested theoretical models for the description of longitudinal profiles in simple alluvial systems. They found quadratic approximations the most flexible descriptor for linked longitudinal profiles.

RECENT TECHNICAL ADVANCES

The combined study of turbulence and sediment transport is everywhere limited by our technical abilities to measure and analyze the active processes. It is not surprising, therefore, that advances in our understanding are often predicated on technical advances. Recent technical advances have improved our ability to characterize bed surface sediments on a local scale, measure sediment transport, and correlate measured velocity signals with images of mixing flows. In order to increase our current capability to characterize the spatial variability of bed sediment caliber, Latullipe *et al.* (2001) developed a visual technique for measuring bed sediment caliber. The technique consisted of a 2-3 day training period whereby operators visually estimated size percentiles and then compared estimates with measurements from established techniques to immediately correct biases. Trained operators were found to be capable of estimating the d_{50} to 15% and the d_{84} to 11%. A calibration relation was found to be independent of individual operators.

Tunnicliffe *et al.* (2000) tackled the problem of high resolution bedload transport by installing a series of magnetic sensors in a stream bed that detect the distortion of the surrounding magnetic field due to the passage of clasts. 82 sensors were used to give a spatial resolution of 10 cm transverse to the flow. Results revealed sediment transport pulses with a period of approximately 20 minutes. Sterling and Church (2002) compared the performance of two more established techniques—Helley-Smith samplers and pit traps—through 22 sampling events in a coarse cobble-gravel bed river. It was found that the two techniques measured different components of the load, with the Helley-Smith samplers collecting substantially less material, as it was not able to trap a large proportion of the coarse load. Rennie *et al.* (2002) developed a non-intrusive technique by using an Acoustic Doppler Profiler (ADP), an instrument more typically applied to measuring flow velocities. Nevertheless, by attributing biases in the bottom-tracking feature of the instrument to bedload velocity, bedload transport was calculated. Results corresponded well with results from conventional techniques, indicating the use of the ADP technique to be suitable for large rivers in an assumed state of carpet flow transport.

The characterization of turbulence in a river is problematic because it is difficult to associate point measurements with specific coherent structures. Roy et al. (1999) present two examples where visual and quantitative techniques are combined in order to overcome this difficulty. The first example took advantage of the natural turbidity contrast between the rivers at the Bayonne-Berthier confluence and filmed the water surface downstream from the confluence in simultaneity with 2 ECMs in the mixing layer and a Helley-Smith bedload sampler. The technique was essential in confirming the role of turbulent structures in mixing and sediment transport. The second example filmed a profile of the river from underwater. For contrast, milk powder was added to the flow behind a pebble cluster. This allowed shedding eddies to be observed simultaneously with flow measurements and allowed the authors to conclude that the shedding process was closely related to the passage of large-scale flow structures.

Latulippe *et al.* (2001) evaluated the feasibility and reliability of visual characterization technique for gravel bed surface material. They showed that the visual technique provides reasonable estimates of the d_{16}, d_{50} and d_{84} of the bed surface. Since Einstein's classical work, tracing methods has become a standard technique in studying sediment transport in gravelbed rivers. Hassan and Ergenzinger (2003) reviewed major available techniques for sediment tracing and developed a conceptual model for their use. The Helley-Smith sampler is the most commonly used, however, the best for obtaining true sediment transport rates in gravelbed rivers is unknown. Sterling and Church (2002) compared the magnitude and texture distribution of sediment samples collected in pit trap and Helley-Smith sampler. The pit trap yielded consistent results with near 100% trapping efficiency for material larger than 2.8 mm. The Helley-Smith was more variable: the sampler exhibit low efficiency for material between 0.71 and 16 mm, and high efficiency for finer material. Automatic detection systems can track the movement of natural and artificial magnetic tracers. Recently, Tunnicliffe *et al.* (2000) developed a new system that provides a high-resolution picture of bedload transport activity.

Stojic *et al.* (1998) used automated DEM acquisition methods to generate dense elevation models to simulate sediment transport in a small physical model (flume) of braided stream. Chandler *et al.* (002) demonstrated the value of the use of terrestrial oblique digital imagery and automated digital photogrammetry for monitoring planform, topography, and rates of change in braided river channels. Wooldridge and Hickin (2002) tested for different methods for the identification of individual step-pool bedforms in Mosquito creek, a small steep stream in British Columbia. The visual identification method provided more information about the bedform geometry than the other methods while the power spectral method identified periodic wavelength in the both examined reaches.

CONCLUDING REMARKS

Fluvial systems in Canada have a number of specific characteristics. First of all, Canada is a high latitude country, which means that, in most basins, the spring snowmelt is a dominant feature of the discharge and sediment transport regimes. Furthermore, in the northern part of the country, the presence of permafrost directly affects hydrological processes and is an important part of understanding fluvial processes and landforms. The northern location of Canada is also important from a historical perspective, since much of Canada was covered by ice during the Quaternary glaciations. As a result, the landscape in most of Canada is relatively young, and rivers are still actively adjusting to deglaciation, which occurred during the late Pleistocene. The high latitude of the country also places it in that part of the world where CGMs generally indicate that the impact of global warming will be greatest in terms of temperature increase. Second, even though Canada is generally viewed as a relatively pristine country, its rivers are rarely unaffected by human activity. A multitude of dams has resulted in modifications of the annual discharge regime, and has led to changes in sediment storage and channel characteristics to a degree that is unknown, but likely substantial. Furthermore, in parts of Canada, human activity has led to a significant degradation of water and sediment quality—typically associated with urban, industrial and, sometimes, agricultural areas—and mobilization of large quantities of sediment within the drainage basin—typically caused by forestry and, sometimes, agriculture. Sediment quality and quantity in a stream directly affect the fish populations and, consequently, studies of the effect of human activity on fish habitat and behaviour form an important, practical part of fluvial geomorphology research in Canada.

Some of the characteristics of Canadian fluvial systems are reflected in the directions of research, for example in the many investigations of bedload transport in gravelbed rivers. In other cases, however, Canadian researchers have not yet taken up the challenge. Relatively little attention, for example, has been paid to the effect of ice on sediment transport and channel morphology, and to the sediment dynamics of Arctic rivers. Overall, Canadian research in fluvial geomorphology during the period of this review (1999-2002) continues in the same direction as earlier work summarized by Ashmore *et al.* (2000). There is a penchant for, and as a result, substantial progress has been made in, investigating the details of fluvial processes at relatively small scales. Examples of this emphasis are the investigations of floc structure, turbulence characteristics and bedload transport, which continue to form central themes in fluvial research in Canada. Translating the knowledge of small-scale, process-related research to an understanding of the behaviour of large-scale fluvial systems, however, continues to be a formidable challenge. Models play a prominent role in elucidating the link between small-scale processes and large-scale fluvial geomorphology, as they do in other fields such as climatology and oceanography. Canadian fluvial geomorphologists have recognized this role of models and, as a result, a number of papers describing models and modelling results have been published during the review period. In the future, it is to be expected that the combination of detailed process measurements and models will gain importance in fluvial geomorphology in Canada, and will lead to an increased understanding of large-scale fluvial systems and strengthen the links between fundamental and applied research.

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Progress in Isotope Tracer Hydrology in Canada

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Abstract
An overview of current research in isotope hydrology, focusing on recent Canadian contributions is discussed under the headings: precipitation networks, hydrograph separation and groundwater studies, river basin hydrology, lake and catchment hydrology, and paleohydrology. Tracer-based techniques, relying primarily on the naturally-occurring environmental isotopes, have been integrated into a range of hydrological and biogeochemical research programmes, as they effectively compliment physical and chemical techniques. A significant geographic focus of Canadian isotope hydrology research has been on the Mackenzie River Basin, forming contributions to programmes such as the Global Energy and Water Cycle Experiment (GEWEX). Canadian research has also directly supported international efforts such as the International Atomic Energy Agency's (IAEA) Global Network for Isotopes in Precipitation (GNIP) and IAEAs Coordinated Research Project on Large River Basins. One significant trend in Canadian research is toward sustained long-term monitoring of precipitation and river discharge to enable better characterization of spatial and temporal variability in isotope signatures and their underlying causes. One fundamental conclusion drawn from previous studies in Canada is that combined use of $\delta^{18}O$ and $\delta^{2}H$ enables the distinction of precipitation variability from evaporation, which offers significant advantages over use of the individual tracers alone. The study of hydrological controls on water chemistry is one emerging research trend that stems from the unique ability to integrate isotope sampling within both water quality and water quantity surveys.

Introduction
Stable and radioisotope tracers have been widely applied in earth systems studies including hydrological and climatological research for their ability to provide a sharper focus on some of the underlying processes that control chemical and physical behaviour of elements and compounds in the natural environment. The ability to study widespread effects has generally made naturally-occurring tracers more useful and more environmentally accepted than artificially-introduced tracers, as well as more transferable to a broader range of biogeochemicial problems. Isotopes of particular interest for hydrological studies include the stable isotopes of water (^{18}O, ^{2}H), which are incorporated within the water molecule ($H_2{}^{18}O$, $^{1}H^{2}H^{16}O$), and exhibit systematic spatial and temporal variations as a result of isotope fractionations that accompany water cycle phase changes and diffusion. Isotope fractionation produces a natural labelling effect within the global water cycle which has been applied to study a wide range of hydrological and climatic processes at the local, regional, and global scales. Anthropogenic nuclides such as tritium (^{3}H), another isotope incorporated in the water molecule ($^{1}H^{3}H^{16}O$), have also proven useful for studying the dynamics of hydrological systems due to its ability to trace precipitation and recharge in the post-1960s era (e.g. Rank et al. 1998). Other radioactive isotopes such as carbon-14 are used for dating groundwater and other relatively old water sources. Solute isotope systems (e.g. carbon, nitrogen, strontium, boron, sulphur, chloride) also provide cabability for labeling solute and pollution sources and in general for the study of hydrological and biogeochemical processes that control water quality (e.g. Hooper and Kelly 2001). An extensive review of the application of isotope tracers to hydrological studies was recently published by Mook (2000). Herein, we focus on reviewing applications of the stable water isotopes (^{1}H, ^{2}H) which are the most universal tracers in hydrological research, and among the most commonly applied in recent studies in Canada.

The use of stable oxygen and hydrogen isotopes as tracers in hydrologic studies has expanded over the past five decades following the initial description of systematic variations in world precipitation (Craig 1961, Dansgaard 1964), development of theory describing isotopic fractionation during evaporation (Craig and Gordon 1965) and testing and validation under a range of field conditions (e.g. Fritz and Fontes 1980, Gat and Gonfiantini 1981, Gat 1996, Clark and Fritz 1997, Kendall and McDonnell 1998; see also Gibson and Prowse 2000)

Isotopic compositions are expressed conventionally as δ-values, representing deviation in per mil (‰) from the isotopic composition of a specified standard, such that $\delta^{2}H$ or $\delta^{18}O = 1000 \cdot ((R_{sample}/R_{standard}) - 1)$, where R refers to the $^{2}H/^{1}H$ or $^{18}O/^{16}O$ ratios in sample and standard, respectively. The most widely used standard in hydrological applications is Vienna Standard Mean Ocean Water (V-SMOW), which approximates the bulk isotopic composition of the present-day global ocean reservoir, and hence has δ^2H and $\delta^{18}O$ values both defined to be exactly 0 ‰. This is a logical datum for hydroclimate studies since evaporation from the oceans is the fundamental source of global atmospheric moisture, which provides the precipitation input for continental water cycling, and the isotopic composition of the oceans is more-or-less invariant on human time scales. Use of the delta scale referenced to V-SMOW also implies that most precipitation and continental waters will have negative values, indicating a lower heavy isotope content compared to the world oceans.

Isotope variations in precipitation are generally characterized by strong linear correlations between ^{18}O and ^2H which reflect mass-dependant partitioning of the water isotopes in the hydrological cycle. This coupling is exemplified by the Global Meteoric Water Line (Craig, 1961), defined as $\delta^2 H = 8\delta^{18} O + 10$, which closely approximates the observed relation between $\delta^{18}O$ and δ^2H in mean annual amount-weighted precipitation world-wide. Variations in the isotopic composition of precipitation reflect climatic processes, including (i) meteorological conditions in the oceanic source area, (ii) rainout mechanisms (i.e. fraction of precipitable water and continental recycling), (iii) air mass mixing and interaction, and (iv) second-order kinetic effects such as those occurring during snow formation and during evaporation from raindrops (Araguás-Araguás et al. 2000). The existence of the MWL is consistent with a conceptual model where global atmospheric moisture arises primarily from a well-mixed source (i.e., the sub-tropic ocean surface) and undergoes progressive rain-out of mass and heavy isotopes during subsequent poleward atmospheric transport (Edwards et al, in press). These effects produce a general shift towards lower heavy isotope content from coastal to inland areas and with increasing latitude. Strong coupling of air-mass vapour content and isotope depletion, often described as a multi-step Rayleigh-type open-system distillation process, is well-illustrated by comparison of the global fields of precipitable moisture and precipitation δ values (e.g., Birks et al., 2002), and is also reflected indirectly in the spatial relations that are observed between precipitation δ values and air temperatures at mid- to high latitudes (Rozanski et al., 1993). The stability of the MWL, which is essentially a long-term isotope climate normal, is confirmed by the fact that it has changed little despite extensive augmentation of the dataset over the past 40 years, including significant contributions from Canadian networks which expanded spatially in the 1980s. From a monitoring perspective, the most notable Canadian contributions include operation of one of the longest continuous time-series records of ^{18}O, ^2H and ^3H in precipitation (Ottawa), and expansion of networks to the Arctic in the 1990s which has extended the range of isotope climate observations to the northern high-latitudes.

Amount-weighted δ^2H and $\delta^{18}O$ values of monthly precipitation received over the year at individual sites also commonly plot in strongly linear clusters in δ^2H-$\delta^{18}O$ space close to the MWL, and best-fit local MWLs drawn through these clusters can provide isotopic input functions for local hydrological studies. Substantial seasonal variability is typical, especially in cold regions, with winter precipitation generally strongly depleted and more variable in heavy-isotope content compared to that received during the summer season. Seasonal changes typically produce shifts along the MWL, which accounts for the extended range of isotope compositions observed in monthly *versus* annual δ values.

Groundwaters generally reflect the isotopic signature of precipitation in the zone of recharge, although individual reservoirs may acquire signatures reflecting their mean residence time and seasonal timing of inputs, as modified by mixing between other sources such as river water or artificial recharge. While transpiration does not generally fractionate the heavy isotope signature of groundwater, evaporation from bare soil, particularly at low moisture contents may produce evaporative isotopic enrichment of groundwater leading to offset below the MWL (see Gat 1996). Evaporation from surface detention storage prior to recharge may likewise produce minor evaporative enrichment. Mixing of recharge with the *insitu* groundwater pool tends to mute seasonal variations in the latter, particularly in deep, well-connected systems, and often enables distinct labelling of individual rainfall events. This labelling effect has been extensively applied to partition streamflow hydrographs into event and pre-event contributions using isotope hydrograph separation. For large river basins, the lag and degree of dampening of seasonal isotope signals in river discharge can be a useful indicator of mean residence time, and effective groundwater reservoir volumes, respectively (Gibson et al. 2002a). Isotopes have also been important tools for the study of glacial and pro-glacial systems (see Stichler and Schotterer 2000), and a wider range of water cycle processes (Gibson and Prowse 2000, see also Gat and Gonfiantini 1981).

Evaporation produces characteristic heavy-isotope enrichment (a.k.a. kinetic fractionation) in surface waters, owing to lower molecular diffusivities in air of water molecules containing the heavy isotope species (Gonfiantini 1986). The isotopic signatures of neighbouring water bodies receiving input of similar isotopic composition typically lie along more-or-less well-defined linear arrays in δ^2H-$\delta^{18}O$ space, termed Local Evaporation Lines (LELs), which deviate from the MWL (slope $\approx$ 8) along slopes that usually range between 4 and 7, depending on local atmospheric conditions during the evaporation season, primarily relative humidity, temperature, and the isotopic composition of ambient moisture. Studies conducted across northern areas of Canada have also shown a systematic steepening of LELs associated with climate shifts and enhanced seasonality towards higher latitudes, although more northerly areas also tend to have lower maximum offsets. Intersection of the LEL with the MWL often provides a useful empirical approximation of the weighted-mean isotopic composition of input waters to a catchment, while displacement of a given lake water along the LEL provides an index of water balance, which can be quantified in terms of evaporation:inflow ratio (E/I) via isotope-mass balance considerations, as discussed under the section Lake and Catchment Hydrology.

Recent Canadian advances in development of techniques and new applications including precipitation networks, hydrograph separation and groundwater studies, river basin hydrology, lake and catchment hydrology, and paleohydrology are reviewed in the following sections.

Precipitation Networks

Collection and analysis of precipitation isotope data in Canada, a long-standing contribution to the IAEA/WMO Global Network for Isotopes in Precipitation (GNIP) database, has shown that the distribution of amount-weighted isotope fields across Canada reflect differences in the dominant meterological regimes (Pacific, Arctic and/or Gulf Stream) associated with each region (Fritz et al. 1987, Moorman et al. 1996). While these efforts were fundamental in providing local hydrological input functions and calibration for paleoclimate archives, there was growing awareness of the significant value of longer-term networks to monitor ongoing and dynamic evolution of the global water cycle. Essentially, these snapshots of the isotope climatology of Canada were limited by the spatial and temporal patchiness of the existing Canadian data. Recent efforts to improve upon this situation include the Canadian Network for Isotopes in Precipitation (CNIP), initiated as a joint venture between university and government researchers, and supervised by a scientific sub-committee of the Canadian Geophysical Union, to provide the spatial and temporal data necessary to examine the sensitivity of isotope fields to changes in circulation patterns, particularly in northern areas where the signal to noise ratio is much lower. The network consists of 18 stations distributed across Canada (spanning almost 40° of latitude and 70° of longitude) collecting weighted monthly precipitation samples. This marks the first time that both the southern and northern regions of the country have been simultaneously sampled. Sampling of the southern stations was initiated in 1997 to supplement an existing informal arctic network (now formally incorporated in CNIP) resulting in a 5 year dataset for the entire country, including a complete El Niño/Southern Oscillation (ENSO) cycle. The arctic subset of the data includes over a decade of sampling and consequently is suitable for evaluating the isotopic expression of the Arctic Oscillation (AO). The sensitivity of isotope-climate signals to modes of interannual variability such as ENSO and AO is of interest because they are a primary cause of interannual climate variability The effects of the ENSO are felt not only near the source in the equatorial Pacific Ocean, but also at higher latitudes, however, the strength, location and timing of climate variations in extratropical areas are less predictable since they are the result of oceanic and atmospheric teleconnections. The strongest climate anomalies are found during the winter following the 1997 El Niño event consistent with a strengthening of the Pacific North American pattern expected during this period (Shabbar and Khandekar 1996, Shabbar et al. 1997, Bonsal and Lawford 1999, Thompson and Wallace 2001). A review of the associated isotope anomalies is presented in Birks (2003).

Regional networks are also under development in Canada. Building on the initiatives of the Canadian Network for Isotopes in Precipitation (CNIP), the recently established Manitoba Network for Isotopes in Precipitation (MNIP) enlists a network of Manitoba schools to collect precipitation. During Phase 1 of MNIP which began in January 2003, twenty schools have been issued precipitation collection stations and some a number of other locations are being considered to expand the network in the near future (Phase 2). Manitoba, situated in the heart of North America, is an ideal place to study the influence of air masses on the isotopic composition of precipitation. The province is affected by a wide variety of air masses, and importantly, is subject to considerable short and long-term atmospheric circulation variability. For example, a particular month can be dominated by deep upper-level troughing in one year (often associated with colder than normal temperatures), but the same month in the following year can be dominated by pronounced upper-level ridging (usually associated with warmer than normal temperatures). Thus, analogues for a wide variety of circulation anomaly scenarios of interest to climatologists, and others, are observed in Manitoba. The dense distribution of MNIP precipitation collection sites supports the spatial resolution (2.8º longitude) necessary to accurately test atmospheric global circulation models and higher resolution regional climate models that incorporate precipitation isotope data. With a dense network of stations from across the province of Manitoba, the MNIP team will be in a position to investigate the influence of atmospheric anomalies (e.g., height and flow anomalies) on the isotopic composition of precipitation. A successful MNIP programme is expected to foster development of similar precipitation isotope networks in other parts of Canada.

Hydrograph separation and groundwater studies

Two and three component isotope hydrograph separations are commonly applied in small scale catchment studies, particularly during rainfall or storm events, to identify the origin, timing, and pathways of surface and subsurface runoff, with the primary objective of evaluating streamflow generation mechanisms. Often isotope tracers are applied in conjunction with geochemical tracers (major ions, trace elements, dissolved organic carbon etc.) which provide capability for labeling interaction with specific substrate materials such as bedrock or organic layers. One of the earliest applications of stable isotopes to define the pre and event water components of watershed runoff was carried out by Fritz et al. (1976), with subsequent applications in various physiographic regions of Canada (e.g., Sklash et al. 1976, Krouse et al. 1978, Sklash 1978, 1979, Sklash and Farvolden 1980, 1982, Bottomley et al. 1985, 1986, Ombradovic and Sklash 1986, Blowes and Gillham 1988, Moore 1989, Wels et al. 1990, 1991a&b, Buttle and Sami 1990, 1992, Gibson et al. 1993a, Waddington et al. 1993, Allan and Roulet 1994, Hinton et al. 1994, Buttle et al. 1995, McLean et al. 1995, Peters et al. 1995). Comprehensive overviews of methodology and field applications have been presented elsewhere (e.g. Kendall and McDonnell 1998, Buttle 1994). The large majority of past studies have established that stormflow in small, forested or wetland headwater catchments in Canada is dominated (>60%) by water stored in the basin prior to a runoff event. A number of recent studies in Canada, conducted in a wide range of hydroclimatic settings have re-affirmed these findings (e.g., Laudon and Slaymaker 1997, Buttle and Peters 1997, Cey et al. 1998, Brassard et al. 2000, Gibson et al. 2000, Metcalfe and Buttle 2001, Fitzgerald et al. in press).

In general, isotope hydrograph separation, applied in conjunction with physical monitoring, has been helpful for establishing or re-defining conceptual models of water delivery on the hillslope or small catchment scale. Buttle and Peters (1997) found that simultaneous monitoring of conservative and non-conservative tracers in streamflow offers additional insight on the age and flow paths of water reaching the basin outlet. The study of Metcalfe and Buttle (2001) also highlighted the need to carry out multi-event or multi-year hydrometric measurements and water sampling in poorly drained boreal landscape dominated by wetlands because they found that the dominance of source waters (old vs. new water) and flow pathways (surface water vs. deep groundwaters) varied from year-to-year depending on the intensity of the snowmelt period, the amount of pre-melt storage of water, and the extent of soil thawing. For instance, a frost table close to the surface and large pre-melt storage in surface depression will lead to high flows primarily of melt water routed over wetland surfaces. Whereas, a year with low melt intensity, low storage levels and greater active layer development was dominated by old water contributions traveling to the stream channel via deeper wetland subsurface routes.

A modelling study of a headwater wetland in Oak Ridge Moraine in southern Ontario by Brassard et al. (2000) revealed that groundwater-surface water mixing during rainstorm events could explain the majority of isotopic signature in the stream and could also be used to measure the extent of potential secondary runoff mechanisms such as pipeflow. A study of the interaction between groundwater and surface water was also undertaken in a small agricultural watershed in southern Ontario (Cey et al. 1998). Hydrograph separations were conducted using $\delta^{18}O$ and electrical conductivity on two large rainfall events with different antecedent moisture conditions in the catchment. Both events showed that pre-event water (groundwater) dominated streamflow and flow in tile drains, with 64-80% of the total discharge contributed by pre-event water. An innovative study by Spoelstra et al. (2001) also applied isotopes of nitrate and oxygen to distinguish between two sources of nitrate in surface waters and groundwaters in two forested catchments within the Turkey Lake watersheds, Ontario. Waddington and Devito (2001) presented a novel approach to hydrograph separation using an inexpensive irrigation device for artificial application of environmental tracers in hillslope and wetland runoff studies.

In a steep bog-forest watershed in north coastal British Columbia, a 3-component hydrograph separation analysis using $\delta^{18}O$ and $\delta^{2}H$ showed that shallow hillslope groundwater dominated bog and event water as runoff generation sources in a hyper-maritime bog-forest catchment (Gibson et al. 2000). Importantly, this study also showed that the deuterium excess parameter, defined as $d = \delta^2 H - 8\delta^{18} O$ (see Froehlich et al. 2002), which varies seasonally in West Coast precipitation, has the potential to label water sources (e.g. bogs, shallow and deep groundwater) according to their residence times. Observations at a nearby coastal headwater swamp revealed that the groundwater regime is dominated by rapid infiltration and short, emergent flow paths. With relatively short turnover time, potential disturbances to the system by harvesting of upslope areas can be expected to occur rapidly (Fitzgerald et al., 2003). At higher elevation sites in the Coast Mountains of British Columbia, isotope hydrograph separation predicted consistent high pre-storm water contribution from subalpline and alpine basin outlets (Laudon and Slaymaker 1997). The authors suggested that pressure propagation from the macropore (fractured bedrock) system could generate the rapid influx of stored water to the stream channel and rainfall was believed to runoff as overland flow due the steep slopes in combination with hydrophobic soils until it can enter the subsurface environment.

Recent advances have also been made to understand the process of isotope fractionation during snowmelt, which significant implications for application of hydrograph separation during spring freshet in the Canadian environment. Detailed laboratory experiments and modelling have shown that a 1-4‰ enrichment in $\delta^{18}O$ of meltwater can arise under a plausible range of conditions due to isotopic exchange between liquid and ice as meltwater percolates down through the snow column (Feng et al. 2002, Taylor et al. 2002). In field studies of deep, slow-melting snowpacks, such changes in the isotopic signature of meltwater have also been verified (e.g. Unnikrisha et al. 2002). The main implication of this work is that the bulk snow composition measured prior to melt may not be adequately characterized as a static end-member for hydrograph separation analysis, although the magnitude of this effect remains to be measured under a wider range of field conditions, including conditions commonly observed in non-alpine, northern regions of Canada (i.e. lighter snowpacks and rapid melt). Importantly, Laudon et al. (2002) present a simple method that greatly improves the separation of event and pre-event water during snowmelt by accounting for both spatial and temporal change in snowmelt isotopic signal and the temporary storage of meltwater in the catchment.

Groundwater studies have increasingly relied on isotope tracers for labeling water sources and interactions in the subsurface. Examples of these applications include the study of groundwater/surface water interaction in a coastal area of the Great Lakes (Huddart et al. 1999) and in fractured bedrock terrain (Oxtobee and Novakowski, 2002), mine and mineral deposit hydrogeology (Sie and Frape 2002, Douglas et al. 2000, Aravena et al. 2003, Harrison et al., 2000), and groundwater as a climate archive (Birks et al. 2000, Remenda and Birks 1999). Numerous studies have also addressed contaminant tracing (Killey et al. 1998) although an exhaustive review of this topic is beyond the scope of this article.

River Basin Hydrology

In addition to small-scale catchment studies, the stable isotopes of water have capability for application in large river basin studies for partitioning relative contributions of flow derived from uniquely labelled geographical sources or distributed components such as direct precipitation runoff, shallow and deep groundwater, and surface waters including lakes and wetlands. Isotopic responses are often complex in large rivers, reflecting the cumulative influence of hydrological processes from precipitation to discharge and including the influence of groundwater, melting glaciers, dams, lakes, karst terrain, evaporation, snow melt events, and tributary mixing. Recent global initiatives such as IAEAs Coordinated Research Project "Isotope tracing of hydrological processes in large river basins" have shown that river discharge signatures provide insight into the basin-integrated hydroclimate forcings on water cycling such as precipitation variability (e.g. changes in condensation temperature, latitude/altitude of precipitation, air mass mixing and recycling, distance from ocean source, and seasonality) and evaporation from the river or contributing sources (Gibson et al. 2002a). The IAEA network, which includes river basins in arctic, temperate and tropical areas, the arid zone, and lowland and alpine drainages, is poised to monitor monthly isotope signals in runoff over the next five years from 22% of the continental land surface, accounting for approximately 33% of the global river discharge. The Canadian contribution to this programme includes water sampling at 22 stations in the Mackenzie Basin and 3 stations in the Ottawa-St. Lawrence river system, and has stimulated interest in development of similar networks in the South Saskatchewan and Coppermine River basins. In addition to collectively improving global capability for isotope hydrology studies and closing the continental isotope mass balance, researchers involved in the project are seeking to improve understanding of linkages between water and nutrient cycling, pollution sources, salinity controls and other water quality issues, as well as hydrological model validation and climate and environmental change detection, particularly where long-term data sets are available.

Sustained Canadian research efforts to study hydrological processes in large river basins have focused mainly on the Mackenzie River Basin within programmes such as the Mackenzie GEWEX study and more recently the IAEA project, although significant hydrological and biogeochemical tracing studies have also been undertaken on the lower Great Lakes drainage basin (Gat et al. 1994, Machavaran and Krishnamurthy 1995, Yang et al. 1996, Helie et al. 2002), the Ottawa River (Telmer and Veizer 2000), the Frazer River (Cameron et al., 1995) and basins outside of Canada (Lee and Veizer, 2003). The Mackenzie River, draining an area of 1.78 million km^2, incorporates a diverse range of geographic source regions, including 8 of the 15 distinct ecoclimatic regions identified in Canada (Ecoregions Working Group 1989). The basin is mountainous in the west and relatively flat-lying in the east with strong north-south climatic gradients, and generally cold, dry climate conditions compared to other large river basins in the world. As a major contributor of freshwater discharge to the Arctic Ocean, the river is also distinct due to the occurrence of several large lakes (Lake Athabasca, Great Slave Lake, Great Bear Lake) which naturally act as flow, sedimentation, and biogeochemical regulators along its main drainage network. Earliest work to define variations in $\delta^{18}O$ the Mackenzie Basin by Hitchon and Krouse (1972) showed systematic variations in discharge from tributaries and sub-basins. Notably, the most depleted isotope signatures (<-20‰ in $\delta^{18}O$) were observed in tributaries of the Western Cordillera, especially the Mackenzie Mountains (min. of -22.9‰ in $\delta^{18}O$), which are characterized by higher altitude precipitation, greater snowfall, and higher runoff/precipitation ratios than other parts of the basin. In shield-dominated areas to the east of Great Slave Lake and Lake Athabasca, and to a lesser extent in the central boreal-taiga plains, tributary runoff was found to be enriched in $\delta^{18}O$ reflecting contributions from lake and wetland evaporation in low-relief areas where rivers traverse extensive string-of-lakes and bog-fen drainage networks. Oxygen-18 values in major tributaries typically ranged between –16 to –14 ‰ in shield areas with peak enrichment observed in wetland dominated drainage of the south-central Boreal Plain (Wabasca R. ~-13.9‰). A synoptic plot showing evolution of $\delta^{18}O$ along the main stem of the Mackenzie-Athabasca River constructed from the data of Hitchon and Krouse (1972) reveals a pattern of regular fluctuations of oxygen-18 from headwaters to mouth due to interaction of tributaries draining both western alpine regions (with depleted isotope signatures) and eastern lowlands (with enriched isotope signatures), overprinted by lake storage effects. In general, lakes serve a regulatory role in the runoff regime by reducing seasonality of discharge and amplitude of isotope variations. The 2-3 ‰ overall enrichment of oxygen-18 from headwaters to mouth, despite the north-flowing drainage network and northeastward decrease in oxygen-18 in precipitation across the region emphasizes the cumulative importance of open-water evaporation losses in the basin water budget (~10%).

Research within the Mackenzie GEWEX study has focused for the last five years on collection of isotopes in discharge and related hydrologic components in the Liard Basin and five wetland-dominated tributary basins ranging from 200 km^2 to 2050km^2 to assess the timing and relative contributions of snowmelt, groundwater and surface water sources to streamflow (Hayashi et al., submitted, St. Amour et al., submitted). In addition to direct assessment of runoff generation mechanisms, the studies have endeavored to develop isotope tracers as diagnostic variables for evaluating and tuning hydrological models. Seasonality in the isotope composition of discharge has been found to be pronounced, with systematic responses observed during the ice-on and ice-off periods due to changes in dominant source waters. A multi-year time-series of $\delta^{18}O$, as shown for the Liard R. basin near the mouth, reveals strong decline in heavy isotope content during spring freshet due to enhanced snowmelt contributions and a steady increase in heavy isotope content during the ice-off period due to inputs of summer precipitation, groundwater and surface water. A substantial decline in heavy isotope content is also noted for the ice-on period, attributed to the increase in relative contribution of deep groundwater *versus* precipitation and surface water sources under ice. Important set-points in the seasonal cycle include ice-off low flow and ice-on low flow conditions which reflect the maximum surface water+precipitation and maximum deep groundwater contributions, respectively (Gibson and Prowse 2002). Similar responses are observed at the basin outlet (Mackenzie R. at Arctic Red River), where peak flow produced by snowmelt typically occurs in April (~Day 150), and coincides with roughly a 2‰ depletion in $\delta^{18}O$ during typical years. Significantly higher depletions during freshet are often observed in smaller tributaries, with similar recessions to higher $\delta^{18}O$ values in summer and late fall. Reduced isotope variability is generally observed during extended winter periods with thick ice cover, and reflect the predominance of groundwater sources (~19.0‰ in $\delta^{18}O$) although the signature of water derived from lake storage is also evident during some years (max of –17.5‰ in $\delta^{18}O$) (Gibson et al. 2003).

Canadian research has also included analysis of winter streamflow and ice fractionation processes under river-ice cover (Gibson and Prowse 1999, 2002). They described a a multi-year isotope sampling survey conducted in the Liard–Mackenzie River Basins, and show systematic isotopic patterns in vertical cores of congelation ice (black ice) obtained from rivers and from numerous tributaries. Gibson and Prowse (2002) attributed these patterns to primary streamflow signals but with isotope offsets close to the equilibrium ice–water fractionation. The results, including comparisons with the isotopic composition of fall and spring streamflow measured directly in water samples, suggest that isotopic shifts during ice-on occur due to gradual changes in the fraction of flow derived from groundwater, surface water and precipitation sources, similar to the larger rivers.

Longer-term water sampling stations within the Mackenzie River basin were established in 2002 to better capture both spatial and temporal variability in isotopic signatures at gauged locations. These efforts are aimed towards partitioning of water from distinct geographical source regions previously identified by Hitchon and Krouse (1972), and to test and calibrate isotope-capable models (Fekete et al., submitted). Importantly, these surveys will include both $\delta^{18}O$ and $\delta^{2}H$ to enable distinct labelling of cumulative evaporation losses. Ongoing studies such as the NSF-funded Pan-Arctic Transport of Nutrients, Organic Matter, and Suspended Sediments PARTNERS are also endevouring to compare the isotope composition of discharge from the Mackenzie River with other major rivers of the northern circumpolar region, in part to understand and partition the freshwater sources and their interaction within the Arctic Ocean (B. Peterson, Marine Biol. Lab, Woods Hole, pers. comm.). Synoptic surveys of isotope composition are also being conducted along the Mackenzie River during 2003 (T. Dick, U. Manitoba, pers. comm.).

One particularly exciting aspect of the river basin work is that monitoring of isotope signatures in discharge can be applied to characterize basin-integrated evaporation and transpiration as separate, coupled fluxes (e.g. Gibson et al. 1993a, Gibson and Edwards 2002). The ability to partition these vapour transfer mechanisms is based the fact that evaporation causes predictable isotopic enrichment whereas plant-mediated transpiration does not. The basic concept for partitioning is outlined below.

In conventionally gauged catchments, water budgets are normally calculated according to

[1] $$P = ET + Q \pm \Delta S$$ [gauged, no isotope sampling]

where P is precipitation, ET is evapotranspiration, Q is discharge, and ΔS is change in storage (often assumed to be zero for long-term periods with stable climate).

Used independently, mean annual isotopic data from a river can be useful for partitioning the fraction of water loss by evaporation from open waters and soils from the contributing catchment area and allows the following partitioning of the water balance

[2] $$P = E + (T + Q) \pm \Delta S$$ [ungauged, isotope sampling]

which is particularly useful where evaporative enrichment is pronounced (i.e. open water or soil evaporation results in a substantial water loss as a percentage of total water losses).

In a situation where both physical and isotope information are used together, it may be possible to carry out a full partitioning of E and T to distinguish the following components:

[3] $$P = E + T + Q \pm \Delta S$$ [gauged, isotope sampling]

As CO_2 uptake during photosynthesis is coupled to the transpiration processes alone, this approach has allowed for better characterization of the net primary productivity as demonstrated for the Ottawa and Mississippi river basins by Telmer and Veizer (1999) and Lee and Veizer (2003), respectively. Although further research is required to ground-truth evaporation and transpiration partitioning for individual terrain types, the distinct labeling of these fluxes has already been demonstrated from tower-based flux-gradient measurements (Brunel *et al.* 1992; Wang and Yakir, 1996).

Lake and Catchment Water Balance

The isotope mass balance approach for estimating water balance parameters has been demonstrated in previous studies of open water bodies (see Dinçer 1968, Gat 1970, 1981, Zuber 1983, Krabbenhoft *et al.* 1990) and was reviewed in detail by Gat (1995). In general, an isotope-mass balance calculation for a typical lake requires measurement or estimation of the isotopic composition of all relevant water balance components. Liquid components, namely inflows, lake volume, and outflows, can often be directly sampled over time and appropriately weighted to provide mean estimates of their isotopic compositions. Due to problems associated with direct sampling of evaporating moisture, the isotopic composition of evaporate is commonly derived indirectly using the linear resistance model of Craig and Gordon (1965), which requires estimates of temperature, relative humidity and the isotopic composition of ambient atmospheric moisture (δ_A). In a situation where all the isotope composition of all water balance components can be characterized, solution of the isotope-mass balance allows for evaporation to be computed as a fraction of the total inflows (E/I).

A series of recent isotope-based studies have developed and refined approaches for application of isotope mass balance methods in seasonal climates of northern Canada. These studies include detailed comparisons of weekly to monthly evaporation in small, well instrumented lakes using non-steady isotope balance methods (Gibson et al. 1996a&b, Gibson et al.1998; Gibson 2002a), regional comparisons of long-term water balance among lakes in Boreal and Arctic areas (Gibson 2001, 2002a, Gibson et al. 2002b, Gibson and Edwards 2002), and application of evaporation pans and cryogenic vapour sampling to characterize isotopic composition of atmospheric moisture near the ground (Gibson et al. 1999). Overall, these studies have shown that application of isotope mass balance using pan-derived atmospheric moisture and laboratory determined values for kinetic fractionation parameters (see Gonfiantini 1986) yields consistent results for short time periods as compared to conventional water balance where evaporation is determined using Bowen ratio and aerodynamic profiling methods.

Various steady-state and non-steady state formulations have been utilized depending on the particular application (Gibson 2002). Due to pronounced seasonal enrichment in lakes during the ice-free period in many parts of Canada, particularly the northern regions, it is often necessary to apply transient models for weekly or monthly studies (Gibson et al. 1996a, 1998) although simplified steady-state models which account for seasonal fluctuations in the components have also been used for comparative analyses on longer time-scales (Gibson et al. 2002b). Evaporation calculations have commonly been performed independently for the isotopes of oxygen and hydrogen to provide a subsequent check on results. In general, it has been possible to simplify the number of required isotopic measurements through development of relationships between the various components. For example, shallow lakes are generally well-mixed during the open water period, and outflows often acquire the isotopic signature of lake water, such that single water samples can be used to characterize δ_L and δ_Q for a given time. In addition, the isotopic composition of combined inflow sources is often close to that of mean annual precipitation, and atmospheric moisture is typically close to isotopic equilibrium with atmospheric moisture during the evaporation season, such that δ_I and δ_A can be evaluated from time-series records of δ_P. Although beyond the scope of this article, the validity of such assumptions have been discussed elsewhere (Gibson 2001).

In stratified lakes, it is often necessary to account separately for epilimnion and hypolimnion volumes and exchanges, provided these have distinct isotopic compositions (Gat 1995). Neglecting stratification can lead to overestimation of the importance of evaporation loss if sampling is conducted during dry, stratified periods and underestimation of evaporation loss if sampling is conducted during wet, stratified periods (Gibson et al. 2002b). Inflow bypass or short-circuiting of the system may also reduce the effective volume of the lake during wet periods or reduce the effective input in the opposite situation. Incomplete lateral mixing within the lake can also a potential source of error when applying isotope mass balance to large lakes. In principle, incomplete mixing can be characterized by spatial and temporal sampling to constrain potential errors to any desired level of precision, although this is not always practical. A simple approximation for systems with similar epilimnion and hypolimnion compositions is to use an average value to represent the undifferentiated lake volume (Gibson et al. 2002b).

Isotope mass balance methods have been applied in conjunction with regional sampling surveys to compare water balance parameters such as throughflow, water residence time and catchment runoff to lakes, thus providing a quantitative basis for examining natural landscape-lake connections and natural/anthropogenic impacts (e.g. forest fire or harvesting) on biogeochemistry of watershed systems (e.g. Gibson et al. 2002b). In combination with physically based estimates of evaporation and precipitation, the method has also been applied to calculate catchment water yield parameters such as runoff/precipitation ratios (R/P). One significant ambiguity of the recent study by Gibson et al. (2002b) was that derived water yields exceeded the available precipitation input in 15% of the headwater lake systems, although an explanation for this was not provided. Closer examination of these results suggests a frequent pattern of excess water yield (R/P>1) in lower elevation catchments that were also typically dominated by fens, as compared to higher elevation catchments that were typically dominated by bogs and tended to have modest water yields (R/P>1). From this perspective, the results may imply that regional groundwater flow derived from outside the topographic catchment area is significant in many of the fen-dominated, lower elevation basins. This explanation is consistent with basic understanding of the origin and water cycle of fens in the region (Halsey and Devito, submitted).

As noted previously, local and regional sampling surveys in northern Canada have revealed a pronounced latitudinal steepening of the slope of local evaporation lines from about 5 to 7 in δ^2H-$\delta^{18}O$ space over the latitude range of 50 to 71°N. The slope of LELs reported from studies at lower latitudes typically range from about 4 to 5.5 (Dincer 1968; Gat 1995). As explained by Gibson (2002b), the elevated slopes observed at high latitudes likely reflect a progressive decoupling of the hydrologic and atmospheric isotope signals with enhanced seasonality. While the isotopic composition of input to the system varies closely with the mean annual δ_P in all systems, the isotopic composition of atmospheric moisture δ_A is strongly biased toward conditions during the evaporation season, which would tend to produce higher LEL slopes (see Gat 1996). A revised conceptual model of isotopic enrichment in seasonal climates suggests that the use of evaporation-flux-weighted parameters significantly improves the reproduceability of the LEL slopes and improves the consistency of the water balance estimates predicted by each tracer, while maintaining the experimental values for the kinetic fractionations for both oxygen and hydrogen (see Vogt 1976, Gonfiantini 1986). An interesting point that is particularly relevant to paleoclimate studies is that temporal changes in seasonality may have altered the slope of the local evaporation in the past. Application of dual oxygen-18 and deuterium tracers to lake sediment archives may therefore be able to trace changes in paleoslope of the evaporation line to provide a basis for examining past seasonality signals. For modern water balance applications, the use of non-weighted atmospheric moisture values, and standard exchange parameters can result in substantial errors in computed long-term values for evaporation to inflow ratios, particularly for strongly seasonal climates where errors may be as high as 50% for low throughflow, high evaporation lakes (Gibson 2002b).

One promising aspect of the isotope balance approach is that it is field-based and can be readily incorporated in water quality surveys in remote or ungauged basins to provide hydrological control for evaluating the water quantity-quality relationships. Recent studies (McEachern et al. 2000, Prepas et al. 2001) have applied the technique to study phosphorous loadings due to forest fire and landscape variables influencing nutrients and phytoplankton communities in Boreal Subarctic and Boreal Plains regions of northern Alberta. Notably, isotope-based methods enabled the latter study to ascertain that mean water residence time (t) was more than 20-fold longer for upland-dominated lakes than for wetland-dominated ones (11 and 0.5 years, respectively, attributable to the deeper lake basins and smaller water yields within the upland-dominated systems. Positive associations ($r = 0.65$) were also noted between lakewater residence time and alkalinity, conductivity, HCO3–, Mg2+, and K+ suggesting the influence of hydrologic setting, specifically lake flushing rates for some but not all dominant ions. The isotope-based estimates of effective drainage basin area eDBA=(R/P·DBA), where DBA is the topographically defined basin area, were also positively related to colour and percentage bog cover (r2 = 0.40 and 0.37, respectively, $P < 0.001$) and negatively related to drainage basin slope and percentage upland cover (r2 = 0.30 and 0.47, respectively, $P < 0.001$). In wetland-dominated lakes, eDBA was found to be a stronger correlate with DOC and TN than DBA. Within the whole data set, eDBA was more strongly associated with lakewater colour than DBA, suggesting a connection with colour-producing wetlands. Colour concentration, being highest in wetland lakes, was correlated with the ratio of isotopically defined effective drainage basin area to lake volume (eDBA/LV, r 2 = 0.63). Overall, the isotope-based indices allowed for better understanding of the relative differences between water balance among the lakes than the physical watershed characteristics alone, which illustrates the added value of incorporating isotopes in such water quality surveys. Several current studies are also incorporating isotope mass balance to the study of lake systems in the Peace-Athabasca Delta region by BB.Wolfe and others and Oil Sands Regional Aquatic Monitoring Program (RAMP), both regional surveys being situated in northern Alberta.

Paleohydrology

Physically-based understanding of stable isotope behavior in modern lake systems, and the occurrence of robust archives of lakewater $\delta^{18}O$ contained in inorganic and organic fractions of lake sediments have also stimulated a wide range of studies into the past hydrology and hydroclimatology of lakes from stratigraphic analysis of lake sediments. These archives include, for example, bulk carbonate and mollusks (Stuiver 1968, 1970, Fritz et al. 1975, Eicher and Siegenthaler 1976), ostracodes (vonGrafenstein et al. 1999), and lake sediment cellulose (Wolfe et al. 2001). Other substrates requiring further study include kerogen and lipids, which hold much promise for providing records of lake water δ^2H (e.g. Krishnamurthy et al. 1995), and $\delta^{18}O$ in biogenic silica (see Labeyrie, 1974; Shemesh et al., 1992, 1995). Recent studies in northern Canada have extensively relied on archives of lake sediment cellulose, often preserved as identifiable algal cells, within zooplankton fecal pellets or as amorphous organic matter (Wolfe et al. 2001), to reconstruct variations in water and carbon balance during the Holocene (McDonald et al. 1993, Wolfe et al. 1996, Edwards et al. 1996). Notably, their results suggest dynamic water balance fluctuations in lakes associated with the advance and retreat of circumpolar treeline. A similar approach was used to reconstruct the history of hydrologic interactions between Hamilton Harbour and Lake Ontario (Wolfe et al. 2000). A review of the application of stable isotopes in high-latitude hydrology and paleohydrology provides additional details on application of these techniques, including case studies of recent paleohydrology studies in the Peace-Athabasca Delta, northern Alberta, and elsewhere (Edwards et al. in press).

Future Directions

To date, research in Canada has contributed significantly to development and refinement of techniques and applications in isotope hydrology that have also contributed broader insight into water cycling processes both within the geographic boundaries of Canada and internationally. Improvement of Canadian infrastructure for isotope-based research, particularly in the last decade, and a growing awareness of the potential value of incorporating isotope tracers as component of broadly based hydrological research programmes has undoubtedly provided fertile ground for significant advances to be made in the future.

One emerging avenue of new international research is the integration of stable isotopes of water into ocean, atmosphere, and land surface hydrological models. These models provide capability for tracking the isotopic composition of reservoirs and fluxes, and simulating the isotopic fractionation processes. As such, they offer a potentially powerful tool for model evaluation and for examining the underlying causes of water cycle variability. To date, isotopes are operationally implemented in several global climate models (e.g. ECHAM-4, GISS, FORSGC AGCM) and new regional climate models with isotope capability are also under development (e.g. REMO). A version of the Swedish HBV hydrological model incorporating stable isotopes has also been tested at the small catchment scale to simulate isographs (isotope hydrographs), and has been used to develop more realistic hydrological parameterization schemes based on a "soft data" optimization procedure (J.J. McDonnell, Oregon State University. pers. comm.). A U.S. DOE-supported water cycle pilot study involving incorporation of isotopes into a mesoscale atmospheric model (MM5) and several land surface models (NCAR-LSM, TOPMODEL, TOPLATS) is also planned over the next ten years. Great potential also exists for application of isotopes in hydrological models at the continental-scale (e.g. Fekete et al., in press). In future, the coupling of continental scale hydrological models with GCMs, including incorporation of evapotranspiration feedbacks to the atmosphere, could potentially improve the realism of the GCMs water and tracer fluxes. Likewise, such a coupled approach could benefit the study of hydrological processes and isotope distribution at the continental scale, and could serve as a model diagnostic used in a similar way to the HBV model run at the small scale. In order to benefit from advancements in new isotope diagnostic models, a commitment must be made to maintain and preferably expand the current scope of water sampling within hydrometric networks in Canada and globally.

A Review of Canadian Remote Sensing and Hydrology, 1999-2003.

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ABSTRACT
Over the years, research and applications of remote sensing in Canadian hydrology embraced a variety of topics. Researcher conducted over the last 5 years emphasised the development of microwave remote sensing, both active and passive. This paper reviews recent (1999-03) remote sensing contributions in hydrology in Canada. Topics include surface water and wetlands detection, soil moisture, snow and snow water equivalent estimates, freshwater ice, permafrost and glaciers as well as distributed hydrological modeling. A very brief description of the theory underlying each application as well as relevant sensors is presented.

Introduction
Remote sensing has been defined as the science and art of obtaining information about an object, area, or phenomenon through the analyses of data acquired by a sensor that is not in direct contact with the target of investigation. Increasingly, Canadian scientists are recognizing that by using remote sensing and GIS in hydrology we have improved our ability to measure or manage spatial, spectral, and temporal information and also our ability to provide data on the state of the earth's surface. The potential of remote sensing for providing information to hydrologists and water resources practitioners has been recognised since the launch of the first ERTS-1 satellite in 1972. Since that time, scientists have developed algorithms to extract hydrological information from remotely sensed data and to develop new, or adapt existing, hydrological methods capable of making efficient use of this new information. The increasing number of satellite and airborne platforms along with advancements in computer and software technology make it possible to evaluate and quantify large numbers of watershed physical characteristics and state variables.
In most applications remote sensing data are used to assess the hydrological state of a basin or region by estimating various hydrologic-state variables (in the liquid, solid or gas phase) and/or hydrologically significant physiographic variables that can influence hydrologic responses or assist in further understanding of the physical processes involved.. The simplest application of remote sensing imagery is to identify items of interest such as snow-covered areas, surface water extent or sediment plumes. Although simple, this type of information is often the only spatial record available describing the extent of a particular water body, ice field or snow region. The second class involves obtain data such as land cover, geological features, or other hydrologic parameters through interpretation and classification of remotely sensed data. It has long been recognized (Rango, 1980) that remote sensing technology is the most cost-effective means of obtaining such data. The third class involves the direct use of digital information to estimate hydrological state parameters directly. This is normally achieved through correlation of known hydrometric data with remotely sensed data or application of electro-optical models. In most remote sensing studies pertaining to hydrology, land use data (class 2) are primarily used in conjunction with hydrologic models. This paper reviews these applications and documents the current state-of-the-science of remote sensing and hydrology in Canada over the last five years. Although significant progress resulting from Canadian research efforts can be demonstrated, very few applications are yet to be considered as operational hydrologic products or as archived products for hydrologic monitoring. A few notable exceptions include flood mapping using SAR data and large scale snow mapping using passive microwave sensors.

The papers that are described below are reflective of the broad thematic and geographic range of the research in the use of remote sensing in Canada. In general they can be categorised into three broad topics of study, namely the remote sensing of surface water and soil moisture, the remote sensing of snow and ice, and the remote sensing of .the land-surface. Globally, hydrologists have been very active creating algorithms to extract hydrological information from remotely sensed data and to develop new, or adapt existing, hydrological methods capable of making efficient use of this new information. Canadian hydrologists have contributed significantly to this body of knowledge. Future advances in hydrological remote sensing and GIS will likely depend heavily on improved technological and scientific capabilities. The Canadian Hydrological community is well-placed to address the future challenges of hydrological science and water management through the use of geo-spatial techniques. Countries like Canada, with large expanses of low populated or remote areas can benefit greatly from some of the observational techniques presented.

Remote sensing of surface water and moisture

As noted by Pietroniro and Prowse (2002), remote sensing of the liquid hydrosphere commonly involves direct measurement of surface or near-surface hydrologic components such as surface water (lentic or lotic), soil moisture and precipitation. Detection of surface water bodies is one of the most straightforward remote sensing procedures since water absorbs most energy in the near- and middle-infrared wavelengths (~0.8 – 10 μm) whereas vegetation and soil have higher reflectance in these wavelengths. Problems can occur when dense aquatic vegetation dominate the signal in water bodies such as wetlands. Furthermore, the method relies on clear sky conditions since atmospheric water vapour in clouds confuses the surface signals. Microwave imagery provides reflectance information and has the distinct advantage of nearly all-weather viewing. Active RADAR sensors such as ERS-1 and 2 (European Remote Sensing Satellite), JERS-1 (Japanese Earth Resources Satellite) and Radarsat-1 (Canadian Space Agency) have all shown potential for estimating open-water boundaries because of the specular reflection of the incident wave and very low return at the operating angles of these satellites (Crevier and Pultz, 1997).

Detection of soil moisture by remote sensing has been another major scientific challenge and has received much attention in Canada over the last five years. Remote sensed reflectance signals are an integrated signature of land cover, topographic and atmospheric effects. Knowledge of electromagnetic energy interaction within the biosphere and the atmosphere is required to separate these components. The utility of real-time soil moisture estimates for the purposes of hydrological or atmospheric modelling is undoubtedly a major scientific challenge. Techniques for data assimilation of this type of information in atmospheric and hydrological models will be of on-going scientific interest.

Soil Moisture

Surface soil moisture represents the boundary interface between the atmospheric and hydrological water balance, and as such represent the most important flux boundary in hydrology. Despite the know importance, it is likely the most difficult variable to quantify because of the spatial variability. Yet, implementation of water management and conservation practices requires quantitative assessment of the soil water status. This is particularly true in agricultural water management projects and operational hydrological modelling such as flood forecasting. Because soil moisture is so spatially variable, in many instances, it is inferred from more easily obtainable conceptual hydrologic variables such as rainfall, runoff and temperature. The Antecedent Precipitation Index (API) or the Palmer Drought Index (PDI) are often cited variables in the literature which try to capture the state of soil moisture ina basin or region. Conventional in situ measurements of soil moisture are costly and provide only point information which may not represent the true state of the moisture profile in the soil. Remote sensing has the advantage of providing an integrated view of a particular region or basin. However, the signal is typically the result of a mixture of land cover, topographic and atmospheric effects. Reasonable knowledge of electromagnetic energy interaction within the biosphere and the atmosphere is required to separate these components. The soil moisture algorithms developed over the last two decades are usually divided into four main groups, and are categorised by the specific bandwidth of the electromagnetic spectrum that the model is concerned with (Pietroniro and Leconte, 2000). These four groups are divided by the following wavelength regions: the reflected visible and infrared, the thermal infrared, active microwave and passive microwave (Colwell, 1983). Canadian efforts have concentrated on the passive and microwave remote sensing of soil moisture, with the use of active senors most heavily studied of the past five years. Historically, the utility of active sensors was well demonstrated by Ulaby (1974) and since that time, the use of SAR sensors has generated considerable interest for both bare soil moisture discrimination (Ulaby and Batlivala, 1976; Bernard *et al.* 1982) and vegetated surfaces (Ulaby *et al.*, 1984; Dobson and Ulaby, 1986).

Because SAR systems allows fine resolution coverage at the watershed scale, and with the launch in 1995 of the C-Band *Radarsat* satellite, several Canadian universities and government agencies have been involved in research relating microwave remote sensing for hydrological and agricultural applications. Near-surface soil moisture (usually 0-5 cm) can be extracted from SAR imagery using statistical, semi-empirical and physically-based models (Boisvert *et al.*, 1996a). Statistical models, for example linear regression models between radar backscatter and soil moisture, are still the mostly used approach (e.g. Pultz *et al.*, 2002; Biftu and Gan, 1999). A major drawback of these approaches is that the models cannot be transposed to other sites. Semi-empirical and theoretical models can be applied to a variety of surfaces; however there are no existing algorithms for the routine determination of soil moisture from single frequency, single polarization radars. Investigations examining the feasibility of extracting soil moisture using polarimetric data and multi-angle data were carried out in anticipation of the Envisat and Radarsat 2 launches. The potential for Radarsat – 2 to provide enhanced opportunities ti extract hydrological variables was noted by van der Sanden *et al.*, 2002. They noted that Radarsat – 2 will provide HH,VV, HV and VH polarization information at the 5 cm waveband. They alos point out that in terms of hydrological applications, soil moisture, snow and wetland detection will make full use of this cross-pole capability. In related studies, Sokol *et al*, 2002 used polarimetric airborne SAR data obtained from the Environment Canada Convair (5 cm wavelength) to investigate the possibility of removing vegetation effects to discriminate differences in soil moisture in pasture fields in eastern Ontario. They found subtle differences in backscatter response for a small range of moisture conditions; however no conclusive relationship could be developed. A major difficulty in extracting soil moisture from SAR is the complex interactions between the radar signal, sensor and target characteristics such as radar incidence angle, vegetation cover, topography and soil surface roughness. Considerable research has been made over the last five years trying to unravel these mechanisms, using scatterometer data, airborne and satellite imagery. Sahebi *et al*, 2002 proposed a multi-polar and multi-angle configuration of C-band satellite SAR imagery for estimating surface roughness. They examined 6 dielectric models, both theoretical and empirical, for a variety of soil moisture and roughness conditions and concluded that the multi-angle approach was better suited than the multi-polar approach to estimate surface roughness.

Although soil moisture is considered the primary influence in biomass changes, Niemann *et al*, 2002 have focused remote sensing efforts on water availability in a forest canopy. The goal of their research was to assess the potential of airborne imagery in identifying moisture variations in the canopy. AVIRIS data were used to relate growth index values (GI) of Douglas fir (*Pseuostuga menziesii*) to reflectance in the visible, NIR and SWIR wavebands. The results indicted that narrow wavelengths bands of 750 nm, 970 nm and 1165 nm were particularly sensitive to the detection of canopy wetness

Leconte *et al.* (1999) used a semi-empirical radar backscatter model with a reference SAR image to infer a soil surface roughness map that was used to estimate soil moisture from other SAR images of the same area. Although attractive, this approach assumes that surface roughness is time invariant, which is not the case for agricultural fields because of tillage practices and erosion. To circumvent that problem, some researchers have focused on pastures as index fields for the determination of soil moisture (Seglenieks *et al.*, 1999; Pultz *et al.,* 1999) , from which soil moisture maps could be produced using geostatistical approaches (Biftu and Gan, 1999), since roughness on those targets can be considered approximately time invariant. Statistical analyses relating radar backscatter (from airborne, ERS-1 and *Radarsat* images) to soil moisture have shown that the best correlations were found using daily means of field average backscatter and soil moisture.

The natural variability of surface soil moisture is not often well understood and complicates the task of deriving suitable soil moisture algorithms from SAR and other remotely sensed data. As SAR is sensitive to surface soil moisture which can vary significantly both spatially and temporally, it is important that measurements be taken over a short period of time for as many fields as possible. Conventional gravimetric measurements being time consuming, other methods, like Time Domain Reflectometry instruments (TDR) and Portable Dielectric Probes (PDP) have been used in the field. Anctil *et al.*, 2002 point out that understanding the variability of soil moisture at various scales determines the optimum size of spatial grids for hydrological models. This of course can have a direct impact on determining soil moisture from spaceborne or airborne platforms. Anctil *et al.*, 2001 found that for organic soils in in southern Quebec, correlation lengths were similar to those of mineral soils and were estimated to be in the order of 100 m.

Canadian research has focused on active SAR data; however, some research also involved the use of scatterometer and passive microwave sensors to retrieve surface soil moisture estimates. Wagner *et al*, 1999 showed the capability of the scattermoter onboard the ERS-1 and ERS-2 satellites fro soil moisture retrieval. An analysis of ERS-1 data over the prairie regions of Canada showed that total water content in the soil profile might be possible with an accuracy of 10% of field capacity during a substantial drying period. This shows strong potential for monitoring soil moisture at the meso and macro scales (footprint > 2500 km^2) and provides encouraging results for future possibilities of such data being assimilated into numerical weather prediction models, or as diagnostic variables for large scale models such as Global Circulation Models (GCM). The potential of passive microwave technologies for global assessments of soil moisture status was recently investigated by Njoku *et a*l, 2003. This study highlights the results of AMSR-E instrument which operates at frequencies between 6.9-89 GHz and which has a resolution of 60 -5 km. Such an instrument holds great promise and qualitative assessments for the northern Great Plains show great sensitivity to surface moisture (Njoku *et al.*, 2003).

Surface Water

The extent, and to a lesser degree the elevation and volume of water bodies is an important parameter for assessing the hydrological state of lakes, reservoirs or wetlands. From the perspective of remote sensing applications, the detection of lakes and large water bodies and to a limited extent, wetlands has always been a relatively simple process and surface water detection is likely the most straightforward remote sensing procedure since water absorbs most energy in the near- and middle-infrared wavelengths (> 0.8μm) and there is little energy available for reflection. The end-result is that delineating surface waters is most easily done using these wavelengths (Swain and Davis, 1978). In contrast, vegetation and soil, have lower reflectance in the visible bands and higher reflectance in the near- and mid-infrared wavelengths resulting in water bodies appearing dark in stark contrast to surrounding vegetative and soil features, particular in the near-infrared wavelengths (Swain and Davis, 1978). Complications can arise when lakes or wetlands are surrounded by significant amounts of submerged vegetation, or are covered with floating vegetation, and often microwave sensors provide additional insight into resolving these problems. Microwave remote sensing platforms are also sensitive to water discrimination and have the distinct advantage of nearly all-weather viewing. Active sensors such as ERS-1 and 2, JERS-1 and Radarsat have all shown potential for estimating open water boundaries because of the specular reflection of the incident wave and very low return at the operating angles of these satellites (Crevier and Pultz, 1997). It should be noted however, that water surfaces, can be subject to Bragg resonance effects where the incident pulse responds to the short ripples or waves. These surface waves generate a backscatter response at the small wavelength of the imaging radar. The vertically polarized transmit and receive signal on the ERS series of satellites is much more sensitive to surface winds and waves than the same frequency C-HH SAR onboard the Radarsat satellite.

Over the last five years, there have been a host of studies using microwave and optical sensors for surface water and wetland detection in Canada. Remote sensing techniques have increasingly been used for wetland assessment and since the 1980's, remotely sensed data from satellites has been considered the most important tool for the identification and monitoring of wetlands (Wang *et al.,* 1998). As noted by Pietroniro and Leconte (2000), much of the wetland remote sensing work in Canada has focused on the Peace-Athabasca Delta (PAD), located at the confluence of the Peace and Athabasca rivers in north eastern Alberta, the Mackenzie delta where the Mackenzie River joins the Beaufort Sea. Furthermore, the Canadian ADRO program was designed to encourage, hydrologic applications for Radarsat and wetland discrimination was seen as a research need as part of that program. Radarsat C-HH data was successfully used (Adam *et al.*, 1998) to map open and flooded vegetation in a the large wetland region within the PAD. This work identified floodwater distribution by segmenting the images into three distinct classes: open water, flooded willow, and non-flooded areas. Töyrä *et al.*, 2001 found that for a similar part of the PAD that was more diverse in terms of vegetation, validation errors using the same technique were in the order of 90% as well. In their study, the synergistic combination of Radarsat microwave information along with SPOT optical imagery were analysed for classifying open water, flooded vegetation and non-flooded areas. Five combinations of image channels were examined for classification accuracy. They were: the *Radarsat* channel alone, the SPOT near infrared channel alone, a combination of all three SPOT channels (green, red, and near infrared), a combination of the *Radarsat* channel and the SPOT near infrared channel and a combination of the *Radarsat* channel with all three SPOT channels. Results clearly demonstrated that Radarsat alone was a better discriminator of these covers than SPOT, but that the combination of all three bands of SPOT with Radarsat provided Kappa coefficients of 99%. A follow-up study (Töyrä *et al.*, 2001b, 2002) looked at the potential to provide a time series of information to track the changes in flooding within this important deltaic ecosystem. Results indicate that with limited ground truth, it was feasible to obtain reasonably accurate flood extent maps during both the leaf-off and leaf–on periods in two years. This time series provides a unique opportunity to examine the changes in the region, and monitor these changes. Such data should be looked upon an important means of obtaining non-traditional hydrological information. This approach was taken a step further by Pietroniro *et al*, 2001, where this data where qualitative assessments between the flood maps derived from satellite imagery, and the estimates of flood extent derived from a one-dimensional hydrodynamic model. Results show that deficiencies in digital elevation information were notable when the imagery was compared with the one-dimensional model. As a result, Pietroniro and Töyrä (2001b, 2002) demonstrated the possibility of using airborne LiDAR (Light Detection And Ranging) information for extraction of digital elevation data. Limited region of the PAD where sampled and results showed good agreement with a consistent bias between ground surveyed points and the LiDAR elevation estimates. However, such new technologies show great promise for hydrological application. Similar approaches for wetland monitoring in Happy Valley/Goose Bay, Newfoundland were reported by Sokol *et al*. (2001). They note that radar remote sensing amplitude images alone were useful for quantifying wetland extent but were not adequate to classify these wetlands. Clearly, the feasibility of wetland classification using active-microwave sensor-amplitude information is difficult, and more information (multi-temporal, multi-band or texture) is required. This was highlighted by Arzandeh and Wang, 2002 who note that single-date radar scenes produce poor wetland classification. However multi-date approaches can yield results comparable with optical classifications. In the absence of multi-date imagery, texture classes such as the grey-level co-occurrence matrix (GLCM) using a variety of window sizes and incidence angles provide significantly improved classification results (Arzendeh and Wang, 2002). Wetland applications Given the potential for northern resources development, particularly in natural gas development slated fro the next 15 years, and with a particular focus on the Mackenzie Delta, the utility of existing, future and historic microwave and optical imagery should prove invaluable during both the design and monitoring phases of these developments.

Another important aspect of surface water remote sensing is flood mapping and the same remote sensing approaches highlighted for water-extent estimates for wetlands and delta environments can be applied to other regions where flooding has occurred. The feasibility of discriminating flood regions with satellites data was highlighted in the early 1970's. Surface water inventories are easily detected at optical wavelengths and near-infrared wavelengths and have been used successfully with the ERTS-1 (Landsat MSS) satellite (Rango and Anderson, 1974) and the VHRR-IR channel of the early NOAA satellites (Weisnet *et al.*, 1974). Active SAR sensors onboard ERS-1 and 2, JERS-1 and Radarsat have all shown potential for estimating open water boundaries (Crevier and Pultz, 1997). Ormsby and Blanchard (1985) carried out some of the earliest experimental work on SAR response to flooded vegetation using X-band, C-band and L-band imagery. They concluded that response depended on wavelength, plant volume and the geometry of the inundated vegetation. In many instances, the combination of vegetation and floodwater provided a bright response due to the "double bounce" phenomenon of the specular reflection on the water and energy returned to the sensor reflected from the vegetation. Active sensors have two distinct advantages over optical sensors for flood delineation. The first is the all-weather capability; in times of serious flood, cloud cover can often be coincident with flood extent. The second is the ability at some wavelengths, to penetrate vegetation. Radarsat and now Envisat have the other advantage of a steerable beam, allowing more frequent coverage of an area of interest.

As noted in a previous review (Pietroniro and Leconte, 2000), Canadian researchers have focused flood line delineation efforts towards Radarsat applications, with the first test of operational flood monitoring occurring in the spring of 1996, 6 months after the launch. As it turns out the 1996 flood event was simply a dry run for the much larger "flood of the century" in the spring of 1997. During the 1997 event, these satellites provided near-daily reconnaissance of flood extent because of this ability to target specific areas. In terms of post-flood analysis, the data is proving useful in assisting in the calibration of a hydrodynamic flood routing model (Saper *et al.*, 1999). As a result of the 1997 Manitoba flood, this relatively simple task of delineating flood extents has evolved into design of prototype flood-information management systems (Wood *et al.*, 2002). This prototype system integrated *in situ* data, remote sensing imagery and web-enabled GIS information in combination with hydraulic and hydrological data to provide useful real-time data and model results. In a similar approach, Deschamps *et al.*, 2002 point out some important issues in dealing with geo-spatial data, particularly as it applies to international drainage basins. They note that major discrepancy between the USA and Canada are: classification systems, datum, feature density variation and attribute discontinuities. As part of the recommended international geo-database, Deschamps *et al*, 2002 describe the utility of flood extent data derived from radarsat as part of the overall geo-database. Pultz and Scofield (2002) also highlight the potential of polar and geo-stationary satellites for hydrological applications. They recommend several key research and development areas to further enhance the use of geo-spatial information in flood hazard management. These recommendations include optimising the use of multi-scale data, research into model parameterisation using remote sensing information, and the use of in situ networks and observation for supporting application (Pultz and Scofield, 2002).

The properties of natural microwave emissions from the earth can also be used to distinguish between open water and lake ice. Walker (1999) has used the Special Sensor Microwave Imager (SSM/I) 85 GHz brightness temperatures to discriminate between areas of ice cover and open water on the same large lakes in the Mackenzie basin in order to assess spatial and temporal patterns of ice freeze-up and decay. Low brightness temperatures over known lake areas indicate the presence of open water (Walker, 1999). Fily *et al.* (2003) developed a simple retrieval method to estimate fractional lake area estimates over the Canadian sub-arctic regions. this methodology requires SSM/I 19 and 37 GHz frequency data to relate the fraction of water covered area to the observed brightness temperatures. Interestingly, Fily *et al.*, (2003) were able to derive time series information for specific “pixel” locations and track the changes in surface water changes over time. Given that the pixel size is approximated to about 25 km, these results may prove quite useful in assimilating this information into numerical weather or hydrological models. As noted by, Bussieres *et al.*, 2002, higher spatial resolution in earth-atmospheric models requires improved information on surface fields highlighting the spatial heterogeneity and thermal response of varying landscapes. Remote sensing is seen as one important avenue to address data gaps in earth-atmospheric models. Recent advances in land-surface interaction, particularly recent finding under the Canadian GEWEX programme (The Mackenzie GEWEX Study or MAGS) on the importance of both small and large lakes to he understanding of the energy balance in the Mackenzie basin have further highlighted this need (Rouse *et al*, 2002). Bussieres *et al*, 2002, conducted experiments with AVHRR data to track changes in surface water temperature in large Boreal lakes of Northern Canada. They found that the temperature cycle over boreal lakes derived from AVHRR measurements fitted a quadratic function (during the summer period) and allowed for interpolation of temperature between cloud free days. As such, they recommend that by using de-trended station observations in conjunction with a quadratic interpolation of surface temperatures derived from AVHRR data will provide a reasonable time-series of average lake temperatures for large northern lakes.

Remote Sensing of Snow and Ice

Snow (and to a lesser degree) permanent ice make up a significant fraction of the available water for consumptive use in many regions of Canada. Brown *et al.*, 2000 document the clear decline in traditional snow monitoring networks in Canada. They also note that remote sensing has been shown to be a particularly valuable tool for obtaining relevant snow data that can be used in, for example, the forecasting of snowmelt runoff in real time and as a useful methods of data extension for long-term climatology estimates. To date, snow can readily be identified and mapped with the visible bands of satellite imagery and use of satellite data to map snow cover extent has become operational in several regions of the world (Pietroniro and Prowse, 2002). For example, digital snow cover maps are produced operationally on a weekly basis in North America NOAA-AVHRR images by the US National Weather Service National Operational Hydrologic Remote Sensing Center.(Carroll, 1995). Finer resolution images such as from Landsat TM (Thematic Mapper) and SPOT-HRV (Systeme Probatoire D'Observation de la Terre – High Resolution Visible Imaging System) have improved the quantitative digital mapping of snow and ice features. Knowledge of temporal and spatial distribution of snow is very important to water resource managers and climate researchers. Snow pack accumulation and ablation, especially during the spring thaw, is a significant input in daily hydrological forecasting systems, which in turn are extremely useful for flood prevention and hydroelectric generation. As Canadian scientists well understand, snowmelt is the most significant hydrological event of the year. In the case of the Boreal ecozone, for instance, the duration and timing of snowmelt exerts strong controls over net primary production and exchange of energy and carbon (Metcalfe and Buttle, 1998). Feedback to the atmosphere is also well understood as extended snow cover delays air temperature warming in the spring (Cohen, 1994), and is can have considerable local, regional and global influence on energy exchange through both a high surface albedo and low thermal conductivity (Walsh, 1987).

Snow

As reported by Pietroniro and Leconte, (2000) research in remote sensing of snow by Canadian scientists mainly focused on active and passive microwave data. This trend has continued, particularly with cryspheric remote sensing programs like CRYSYS. Significant advances were made to elucidate the complex interactions between microwave signatures and snow cover properties, such as areal extent, SWE, snow depth, albedo and grain size. Microwave remote sensing offers great promise for applications to snow hydrology. Since the early 1980s, the Meterological Service of Canada (MSC) has developed expertise on the use of passive microwaves for estimating Snow Water Equivalent (SWE) for dry snow. Maps of SWE for the Canadian Prairies are produced on a weekly basis using SSM/I data and distributed to operational hydrological forecasters (Goodison *et al.*, 1990). Poor spatial resolution (25 km) limits the use of SSM/I to very large areas. Active microwave sensors, such as SAR, can discriminate between wet and dry snow (Shi *et al.*, 1994) and have spatial resolutions (10-100 m) making possible hydrological studies at the watershed scale.

Using SSM/I data, Goodison and Walker (1995) have developed a method for real-time estimation of areal snow water equivalent (SWE) over the Canadian Prairies using a vertically polarized brightness temperature gradient (VTG). A linear regression model between VTG and field data has shown a correlation of 0.89 for dry snow areas. As part of the MAGS programme, Walker and Silis (2002) produced a time series of SSM/I derived SWE for March 1 of each year between 1988 and 1998. Because of the heterogeneity of the landscape in the Mackenzie basin, th authors used the IGBP land-cover classification, aggregated to 4 land-clasees and applied a weighted algorithm to the SSM/I pixel in order to obtain the SWE estimate. Walker and Silis (2002) conclude that algorithms developed for boreal and prairie landscapes for SSM/I appear to perform well in the Mackenzie River basin. These maps are also providing an important tool for assessing Regional Climate Model (RCM) performance as part of the MAGS science plan. Winter precipitation for specific years derived using the Canadian RCM linked with the CLASS-LSS are being compared with these remote sensing outputs (Walker and Silis, 2002).

Given the now relatively long time series of prairie snow-cover imagery, Derksen et al., 2000 used a series of late spring imagery and principal component analysis to examine spatial patterns in prairie snow distributions. Using 5-day composites derived from SSM/I regressions, they compared the inter-annual variation between the SWE maps and found that snow cover fell into tow distinct regimes during the available study period between 1988 and 1995. Attempts at relating these changes in spatial patterns to synoptic climatology are underway, and may provide important insight into the climatological behaviour of snow on the Canadian prairies. In follow-up work, Derrksen et al, 2002 assessed the acuracy of SSM/I derived SWE maps for the Canadina prairies and detrmined that there was no general bias in prairie-grassland SWE estimates, however systematic underestimation was a problem in dense conifer forests (Derksen *et al*, 2002).

These SSM/I applications have been extended recently and are compared to the MODIS operational SWE maps for the Canadian Prairies. Bussieres *et al.*, 2002 evaluated the MODIS-MOD10 snow extent maps for the Canadian Prairies and the northern boreal forests of Quebec. They show that the MODIS product has good agreement with SSMI derive SWE estimates and provides SWE at a much higher resolution (1 km as opposed to 25 km with SSM/I). Similar work by Hall *et al.*, 2002, looked at the relative accuracy of hemispherical-scale now-cover maps and also compared SSM/I and MODIS derived snow-cover maps. Results show large similarities between MODIS derived maps (which are based on visible and near-infrared algorithms) and the SMM/I derived snow maps. Hall *et al.*, 2002 also point out that with the launch of the NASA Aqua satellite in 2001, snow mapping will be developed using the AMSR-E microwave scanning sensor with an improved resolution of 12.5 km, and allow for better comparisons with visible based MODIS data.

Although most studies are limited to examining results from SSM/I or MODIS, other researchers have been experimenting with different frequencies of both passive and active microwave sensors to assess snow cover remotely. Nghiem and Tsai (2001) present interesting results derived from Ku-band scatterometer data, examining the SWE condition preceding the 1997 “flood of the Century” in Manitoba. They note that Ku-band backscatter is quite sensitive to changing snow conditions and in particular snow properties. Using NSCAT data, Nghiem and Tsai (2001) show that global backscatter signatures in snow-covered region reveals patterns consistent to global snow-classification systems, and moreover, comparison with ground observations of snow-depth and temperature data are well correlated.

New retrieval algorithms from passive microwave data have also been conducted and based largely on snow emission models. One example was to evaluate the Helsinki University of Technology (HUT) emission model in the BOREAS study area (Roy *et al.*, 2002). In this study, they demonstrate some promise of better SWE retrieval from the boreal forest over empirical relationships already established. Roy et al, 2002 show that by inverting the model using a minimisation technique, SWE estimate were within 15 mm with no appreciable bias.

Active remote sensing of snow has also been the focus of many investigations over the last 5 years. The determination of SWE using SAR is a difficult task because dry snow is virtually transparent at radar frequencies such as those onboard aircrafts and satellites. On the other hand, the penetration depth of a microwave signal in even slightly wet snow (<1% wetness) decreases dramatically, preventing extraction of information over the depth of the snowpack.

Because snow is an excellent insulator, it was reasoned that one could indirectly obtain SWE by detecting the presence of unfrozen water into the upper frozen soil layer. Following this reasoning, Bernier and Fortin (1998) obtained a unique relationship between the thermal resistance of the pack and the backscattering power ratio between an airborne SAR winter image and a reference image with no snow. The relationship, applicable to shallow snow packs (SWE < 200 mm) was used to extract snow thermal resistance for a SAR image, from which the SWE was derived using a linear relationship obtained from in-situ observations. Similar relationships were also obtained with ERS-1 data (Bernier et al., 1995). The algorithms developed have been used quasi-operationally in a mapinfo-based application named EQeau which largely based on the work presented by Bernier *et al.*, (1995). The feasibility of extracting SWE using EQeau for the 1998-99 snow season was investigated for eastern Canada. ScanSAR Radarsat images were used to obtain estimates of regional basin-wide SWE and results were within a few millimetres of measured SWE (Bernier et al, 1999). Fortin *et al.*, 2000 go on to show that EQeau was successfully integrated within the Hydro-Quebec water-supply forecasting system. Other SAR-SWE related studies include Leconte *et al.* (1999), who explored the potential of extracting snow depth from RADARSAT C-HH SAR images acquired over agricultural terrain. The approach was based on sensing the soil surface temperature and used to obtain snow depth for dry snow packs from the radar signal corrected to account for the presence of a thick ice at the base of the pack. Although reasonable snow depth estimates were obtained for half of the fields analyzed, it was concluded that mapping the snow depth at the watershed level is still an elusive goal due to the complexity of the interactions between the radar signal and target characteristics. Research on mapping wet snow using airborne and satellite C-band SAR has been ongoing at University of Waterloo since 1991 (Seglenieks *et al.*, 1999. Based on ground truth reference maps created using oblique aerial photography of agricultural terrain within a day of the SAR image acquisition, SAR imagery was classified into snow covered and bare ground. Classification accuracy generally ranged from 60 to more than 80% depending on sensor configuration (incidence angle, resolution, polarization). The classification had difficulty in distinguishing between field with ponded water and wet melting snow, however the total snow covered area estimates were reasonably close to the ground reference maps.

Terrestrial Ice

It has long been well understood that the future of Canadian water resources, environmental quality, and sustainable development require an understanding of glacial and hydrological processes, including the ability to effectively monitor and forecast these resources. Glaciated areas are of considerable importance in Western Canada since they supply melt water for hydro-electricity, irrigation, industry, domestic use, and the development and maintenance of stream associated habitat. Alpine snow and ice mapping at various spectral and temporal frequencies is well documented in the literature (e.g. Dozier, 1984; Shi and Dozier, 1993; Shi *et al.*, 1994, Hall, 1995). Canadian hydrologist and glaciologists have historically viewed remote sensing as an important tool in monitoring glacier fluctuations (Østrem, 1975).

Current remote sensing efforts in Canada in glacier monitoring over the last five years have focused on estimating glacier extent, snowline detection and movement as both a hydrological modelling input as well as a surrogate for mass balance (Demuth and Pietroniro, 1999; Cogley et al, 2001). In this context of glacier monitoring and modelling the snowline extent is often a desired parameter for many applications (Demuth and Pietroniro, 1999). This migrating snow line plays an important role in the energy balance of a glacier basin, controlling the timing and quantity of melt. As noted by Demuth and Pietroniro (1999), during late summer the low albedo of glacier ice and firn causes two to three times more energy absorption at the surface than if it were snow covered (Grenfell and Maykut, 1977; Grenfell and Perovich 1981; Brugman, 1991). Therefore, as the snow line position strongly affects the amount of seasonal melt discharge, it is necessary to map its location for physically based hydrological modelling in alpine areas.

Glacier extent mapping for inventory purposes has also been recognised as an important initiative where remote sensing data can play an important role. As noted by Sidjak and Wheate (1999), glacier inventory mapping in Canada has been based on photogrammetric techniques and air-photo interpretation, and a single Landsat scene captures an area covered by hundreds of aerial photographs, minimises relief displacement and provides sufficient detail to discriminate mots glacial features. Using principal componenet analysis, ratioa along with a normalized difference snow index (NDSI) derived by Hall et al, 1995, Sidjak and Wheate (1999) show that these derived products redue the topographic influences and provide better discrimination of the glacier extent than standard TM bands.

In tandem with the research using optical sensors, Canadian scientists have also focused on SAR applications as well. Snowline detection using SAR imagery is accomplished by minimising the topographically induced radiometric and geometric distortions inherent in SAR imagery and then delineating the wet snow, glacier ice, and bedrock surfaces (Adam *et al.*, 1997a). Because dry snow is transparent at C-band, this method is only effective in mapping the snow line if the snow is wet or actively melting, which typically occurs in the late fall for temperate zone glaciers. Investigations into tracking this snowline movement with satellite SAR are on-going and show some promise (Demuth and Pietroniro, 1999). The late summer snowline extent for temperate glaciers also corresponds to the Equilibrium Line Altitude (*ELA*) and a related parameter, the Accumulation Area Ratio (*AAR*). These parameters can be used to estimate, in the absence of more complete information, the whole-glacier net mass balance of a glacier (Østrem, 1973) may provide information on the regional state of glaciers. In terms of mass balance estimates, Demuth and Pietroniro (1999) clearly showed that with late ablation season imagery, Radarsat SAR data can discriminate, using manual or automated image analysis techniques, ripened first-year snow/firnpack and bare glacier ice. The remote sensing derived snowline and the resulting surrogate glacier mass budget estimates required serious consideration as a tool in the absence of traditional ground measurements used in the direct glaciological method. They concluded, however, that such estimates can vary substantially depending on how representative the applied DEM is of the imaged glacier topography and that these data were still not sufficient replacement for traditional mass balance networks.

In order to extend the mass-balance record for Arctic glaciers, Cogley *et al.*, 2001 have developed a system that makes use of Radarsat "Browse" images. These images are radar (Radarsat) images that have an effective resolution of 2 km and number about 100 scenes per year (Cogley *et al.*, 2001). In the analysis, the authors determine that rather than focus on the equilibrium line location, the time series information derived from these relatively inexpensive browse images contains useful detail on the intensity of melting. They also suggest that there is persuasive evidence that radar monitoring holds promise for estimating glacier melt rates (Cogley *et al.*, 2001).

Another important aspect of terrestrial ice is the duration and depth of lake ice-cover. Lakes are an important part of many ecosystems and can occupy a significant fraction of the total land-system (Goodison *et al*, 1999). Lake-ice has a strong influence on local energy and water exchanges, and lake ice-breakup dates are also good indicatores of regional climate variability (Duguay et al, 2002). Extending the cryospheric mapping focus to freshwater ice, Dugay et al. (2001) demonstrate that Radarsat and ERS-1 can be used to monitor the dates of freeze-up and break-up on lakes. They note that such dates may be good proxy indicators of regional climate change and hence a useful variable to include in the Global Cryosphere Operating System (GCOS). Other interesting applications were recently published by Duguay et al, 2003 and focused on using multi-sensor data to establish lake depth and ice-thickness of shallow sub-arctic lakes.

Land Surface

The land surface is an important control for both the water and energy balance of the earth's surface as it is the primary influence in the surface water budget and it is almost always a required input into both hydrologic and atmospheric models. Over the number of years there has been a proliferation of sophisticated land-surface water and energy balance schemes being implemented in global climate models (GCMs), regional climate models (RCMs) and day-to-day operational forecasting numerical weather prediction models (NWPs). A strong research interest has also developed in coupling atmospheric and hydrologic models within a common framework to achieve more accurate flow predictions. To permit proper partitioning of water and energy over the land, some conceptualisation of the land-surface is also a requisite component of these models. In many of these models, remote sensing offers the best method for obtaining such information.

Land-cover classification is actually one of the earliest satellite-derived products used in hydrological analysis and is often used as a classifier for parameters of a hydrological model (Kite, 1989). Detailed land-surface information is especially important for distributed hydrologic models where the hydrologic response of spatial units is controlled by the nature of the land-cover and vegetation. The distributions of land cover types may be determined by classifying data from any optical satellite imagery and to some degree, microwave imagery. Notably, however, quite different results have been achieved using different forms of imagery, largely due to differences in spectral resolution and/or spatial resolution. Pietroniro *et al.* (2000), for example, demonstrated that large differences exist between land-cover maps of the Mackenzie River basin derived from the relatively coarse NOAA's (National Oceanic and Atmospheric Administration)–AVHRR imagery and the higher resolution Landsat-TM products. Despite such problems, incorporation of satellite-derived land-cover classifications has continued to improve the capability of hydrological models. Most hydrological models and Land-surface schemes (LSS) such as the Canadian Land Surface Scheme (CLASS) (Verseghy *et al.*, 1993) require land-cover information and Leaf Area Index (LAI) for assessing surface fluxes. Albedo estimates for radiation balance, surface roughness length for evapotranspiration and moisture storage in the canopy all require some form of generalized land-cover information (Verseghy *et al.*, 1993). Also, as noted by Pomeroy *et al.*, 1998, both horizontal and vertical fluxes of snow are strongly dependant on the land surface type. Integrarting these products with existing or new hydrological models is a continuing challenge in hydrology. Galarneau *et al*, 2001 proposed the use of soil moisture derived from Radarsat imagery of the Chateauguay river basin in Southern Quebec, in conjcuntion with Richard's equation to establish basin-scale soil moisture maps for use in hydrologcial appliaction. Gan and Biftu (1999) applied a semi-distributed hydrologic model, DPHM-RS, to the Paddle River Basin (265 km2) of Central Alberta. Using the Digital Terrain Elevation Data (DTED), the watershed was divided into five sub-basins drained by a network of tributaries and the main stream. Remotely sensed, spatial information such as leaf area index, albedo, surface temperature, near-surface soil moisture, and land use were used in the model. Simulated runoff at the basin outlet showed good agreement for both calibration and validation stages (R^2 of 85% in 1996 and 60% in 1997).

Conclusions

Many of the existing techniques for using remotely sensed information in hydrology have been described, some of which are in use by operational agencies, and others which are still very experimental. With the recent launch of the of the Earth Observing System (EOS)as well as Envisat, and the potential launch of the Radarsat-2, remote sensing will play an increasingly important role in both monitoring changes in the hydrosphere and cryosphere as well as provide important real-time data for water management purposes.

Efforts should be made by the data suppliers and research agencies to educate potential users in the advantages of remotely sensed data. Certainly, the real strength of satellite remote sensing is in alleviating some of the hydrometric data collection and management problems facing Canada given the large expanses of low populated or remote areas. As Schultz (1988) pointed out, satellite sensors do not measure hydrological data directly; the hydrology is obtained only after interpretation of the measured electromagnetic radiation. In many cases the analysis of remotely sensed data consists only of developing a regression equation between the desired information and the available pixel intensities. As noted by Kite and Pietroniro (1996), such regressions are typically not applicable beyond the time and space constraints of the original data and add little if anything to our understanding of what is actually being measured. If remote sensing is truly to be effective in water management and hydrologic applications, then advancement in the understanding of the electro-magnetic interaction with the hydrosphere and crysophere is critical to the advancement of hydrologic sciences in Canada.

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