Journal of Environmental Informatics 17(1) 36-45 (2011)

Journal of Environmental Informatics

www.iseis.org/jei

Field Investigation and Hydrological Modelling of a Subarctic Wetland - the Deer River Watershed

L. Jing[2], and B. Chen[1,2]*

[1]*Research Academy of Energy and Environmental Studies, North China Electric Power University, Beijing 102206, China*
[2]*Faculty of Engineering and Applied Science, Memorial University of Newfoundland, St. John's, NL A1B 3X5, Canada*

Received 9 May 2010; revised 25 October 2010; accepted 15 November 2010; published online 12 March 2011

ABSTRACT. Recently, investigation and conservation of subarctic wetlands has been recognized as an attractive route. To gain insight of the interactions between hydrology and atmosphere of the second largest wetland in Canada - the Hudson Bay Lowlands (HBL), the semi-distributed land use-based runoff process (SLURP) hydrological model was applied to a typical subarctic wetland - the Deer River watershed over a 20-year period (1978-1997). Sensitivity analysis, calibration and validation of the model identified a number of distinguishable hydrological features of subarctic wetlands as well as model deficiencies. Snowmelt was the major source of water recharge in subarctic wetlands and constituted approximately half of the average annual runoff in the Deer River watershed. The peaks of the simulated spring runoff were 34% lower than the observed ones in average which could be attributed to the effects of shallow permafrost that impeded the infiltration of melt water. Runoff of rainfall water during the summer season occurred only during storms due to canopy interception, depression storage, soil porosity, impermeable permafrost, and intensive evapotranspiration. A lag of 2-8 days between the peaks of streamflow and rainfall was observed through both field investigation and modeling results. The numerous seasonally connected ponds/lakes stretching over the middle and lower reach of the watershed behaved as buffers and significantly prolonged the concentration time in summer and fall. The findings will help build a scientific basis for advancing the knowledge of the hydrologic cycle and impacts of climatic changes on sub-arctic wetlands.

Keywords: hydrological modelling, permafrost, rainfall, subarctic wetland, snowmelt

1. Introduction

Wetlands comprise 14% of Canadian landscape and exist as bogs, fens, swamps, marshes, and shallow water (Price and Waddington, 2000). Their considerable impacts on water storage and distribution, water quality, carbon and nitrogen cycles, regional climate, and ecosystems have been noticed (Price et al., 2005). Recently, public recognition of their environmental significance has highlighted urgent need for in-depth understanding of the hydrological processes in order to more efficiently conserve wetlands and assess climate-related impacts, especially in the northern regions (Rouse et al., 1997; Woo and Young, 2006; Ström and Christensen, 2007; Jing et al., 2009; Li et al., 2010). Arctic and subarctic regions are sensitive to climatic changes and therefore are of crucial importance in maintaining the integration of global environment as well as the arctic communities. A recent report from the Arctic Council and the International Arctic Science Committee (ACIASC) has demonstrated that the temperature of the arctic regions has been elevating sharply at twice the average rate of other regions in the world; moreover, precipitation has been increasing as well at a rate of 8% annually (ACIASC, 2004). Characterizing the hydrologic features of subarctic wetlands is crucial for the purposes of modelling and predicting how the water cycle may vary in present and future.

The hydrological processes of Canadian subarctic wetlands are influenced by a number of factors, including the climatic conditions, the low-relief terrain, and some unique land features, such as seasonal ponds, muskeg, and lichen. Adequate water replenishment from snowfall and rainfall is the primary condition of maintaining the existence of subarctic wetlands. Long cold winters and short mild summers dominate the local macroclimate. Snow accumulates on the ground and vegetation canopy in winter while streamflow depletes and is largely sustained by deep groundwater. Snowmelt in May and June is the major source of water recharge and constitutes approximately half of the annual water input. Although permafrost may impede the infiltration process, water from snowmelt and the thaw of soil ice content penetrates the ground and subsequenttly replenishes surface runoff and groundwater storage. However, this replenishment may be depleted during the summer when evapotranspiration exceeds rainfall amount, causing the streamflow to subside. Water presents in the surface soil layers starts to freeze in fall while groundwater level continues to decrease.

* Corresponding author. Tel.: +1 709 8648958; fax: +1 709 8644042.
E-mail address: bchen@mun.ca (B. Chen).

ISSN: 1726-2135 print/1684-8799 online
 doi:10.3808/jei.201100185

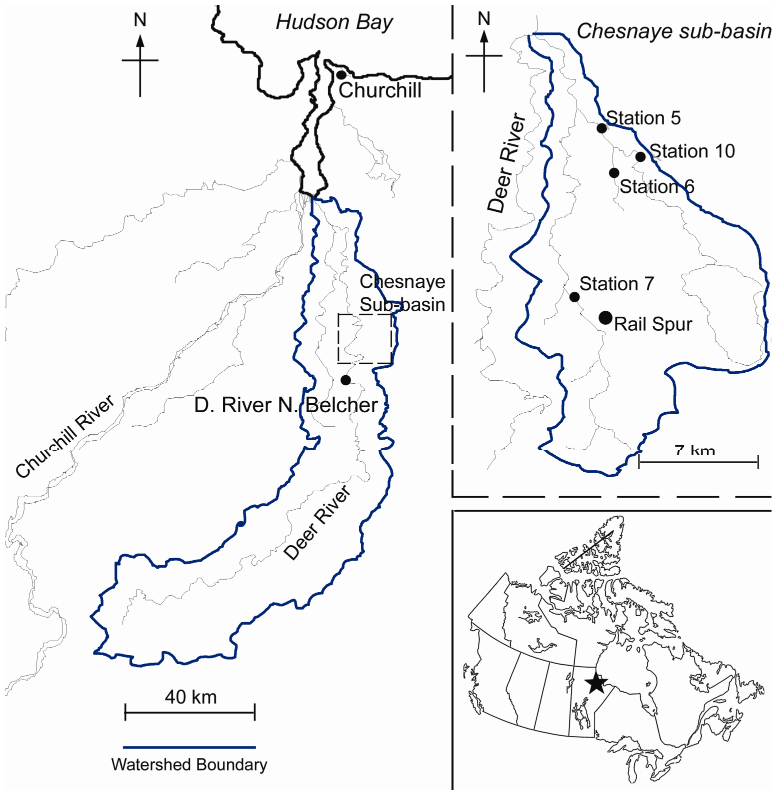

Figure 1. Location of the Deer River watershed and the Chesnaye sub-basin (with monitoring stations).

A multitude of previous studies have explored the characteristics of subarctic wetlands. Sufficient water supplement, which comprise of snowmelt, precipitation, groundwater, streamflow, and inundation from lakes, is the determinant factor of the existence of wetlands (Woo and Young, 2006). Woo and Marsh (2005) reported two distinguished flow mechanisms that occur in the hummocky terrain and organic-mineral layers systems. Channel runoff resulting from snowmelt and rainfall is mainly delayed by lakes in the vicinity and the particular permafrost (Quinton and Roulet, 1998; Leenders and Woo, 2002). Soil features of subarctic wetlands play a key role in hydrological processes because the porosity and hydraulic conductivity dramatically declines with depth (Woo and Marsh, 2005). Recently, variation in climatic conditions, which couples with changes in the magnitude of water supply, permafrost degradation, and even complete drying, has attracted much attention (Payette et al., 2001; Woo and Young, 2006). The temperature of the subarctic regions, especially the Hudson Bay Lowlands (HBL) which is the second largest wetland in Canada, has been increasing during the past decades (Rouse et al., 1997).

To help better understand and predict the water cycle, hydrological modelling has been playing a significant role in studying the distinguished attributes of subarctic wetlands. Numerous models have been developed and applied for hydrology simulation, such as hydrologic engineering center (HEC-1), semi-distributed land use-based runoff process (SLURP), precipitation-runoff modelling system (PRMS), streamflow synthesis and reservoir regulation (SSARR), snowmelt runoff model (SRM), UBC, WATFLOOD, and TOPMODEL (Quick and Pipes, 1977; Beven et al., 1984; Abbott et al., 1986; Bergström, 1992; Kouwen et al., 1993; Bicknell et al., 1997; Kite, 1997; Richard and Gratton, 2001; Chen et al., 2003, 2004, 2007, 2008). However, only a few studies specifically targeted at subarctic wetlands, particularly the HBL in northern Manitoba due to a number of knowledge gaps, such as the difficulty of considering continuous permafrost, numerous seasonal ponds, and snow sublimation as well as the complexity of simulating water linkage between surface and subsurface flows in the hummocky terrain (Mancell et al., 2000; Zhang et al., 2000; Van der Linden and Woo, 2003; Boswell and Olyphant, 2007). As one of the most remarkable attributes, the overlaying permafrost existing in northern Canada seasonally thaws during the summer and refreezes during the cold long winter. Its depth varies by season and location which significantly influences the generation of interflow and groundwater flow. The presence of numerous ponds and hummocky terrain also poses difficulties in recognizing the behaviour of water transfer between surface and subsurface flows due to the vast and uncertain water storage capacities of ponds and organic soil layers. Field and modelling efforts have been limited on these gaps and their influence on hydrological modelling regimes has not been well studied. In recent decades, research and development of advanced modelling approaches have been driven by the urgent needs in managing and conserving subarctic wetlands under changing climatic conditions. Among many models, the SLURP model has been originally developed for meso- and macroscale basins with intermediate complexity, which incorporates necessary physical processes without compromising the simplicity of calculation. It has been widely used in Canada and other places in the world because of the availability of source code and relatively low parameter requirement. However, few studies have been reported to apply the SLUPR model to Canadian subarctic wetlands especially the HBL due to the above knowledge gaps. Therefore, the objective of this study is to describe an application of the SLURP model to subarctic wetlands in the HBL and then evaluate the performance of simulating each hydrological process as well as identify the advantages and disadvantages of using the SLURP model in such areas. This study will not only help understand the interactions between climate and hydrological processes in subarctic wetlands, but also guide others attempting to improve their simulation efficiency in regions of hydrological response.

2. The Study Area

The Deer River watershed (57°55' N, 94°46' W) is located in the northern part of the HBL and 70 km south of the town of Churchill, Manitoba (Figure 1). It lies on the edge of high subarctic and low subarctic regions. The drainage area is 5,048 km^2 with elevation gradually descending from 232 m in the southwest to 16 m in the northeast (Jing et al., 2009). The Deer River, as one of the largest tributary of the Churchill River, originates in the southeast region of the watershed and flows towards northeast. The watershed is a broad polygonized peat plateau which consists of high and low centered polygons. Vegetation is predominantly lichen-heath, in which lichen coverage varies with sub-basins and ranges from 67 to 83% in the watershed. Deciduous and evergreen forests dominate the headwaters of the Deer River and adjacent areas of downstream channels. The hummocky terrain prevails and consists of porous peat overlying a thick layer of mineral substrate. Continuous permafrost underlies the study area with an active depth of approximately 1 m by late August. The primary soils are brunisolic static cryosol, brunisols, brunisolic turbic cryosol, and organo cryosol according to Mills et al. (1976). Reeve et al. (2000) collected data across the HBL and reported that hydraulic conductivity drops from 4.2×10^{-6} m/s at a depth of 1 m to 1.5×10^{-6} m/s at a depth of 2 m. Wessel and Rouse (1994) stated that volumetric water content of peat soil and hummock is capable of reaching at 80 to 90% and 50 to 60%, respectively. Subsurface water content reaches its maximum equilibrium following final snowmelt when surface water fully recharges the soil layers.

A representative sub-basin in the lower reach of the Deer River, the Chesnaye sub-basin, was selected for the extensive field investigation during 2006 ~ 2008 (Figure 1). It is extremely flat with elevation slightly varying around 52 m. Vegetation is mainly tundra and shrub with sparse coniferous forest along the streams. Many seasonally connected lakes and ponds stretch over the basin. The Canada VIA railway goes through the basin from north to south and makes the Chesnaye sub-basin accessible. A monitoring network of four stream gauging stations (i.e., Stations 5, 6, 7, and 10) and one automated weather station (i.e., Rail Spur) was maintained during 2006 ~ 2008 for supporting the investigation and modeling work of this 130 km^2 sub-basin (Jing, 2009).

The Deer River watershed has a marine subarctic climate. Winters are long and cold with average temperature varying around -20 ^{o}C from November to April inclusive. Snowmelt commences in early or mid May and snow has typically disappeared by the beginning of June. Summers are cool and short in which all plant growth occurs. Soil water deficits are common in summer and fall due to the intensive evapotranspiration. The records show that the nearest town, Churchill, experienced a mean annual temperature of -6.6 ^{o}C during 1978 ~ 2008 (Table 1). Mean annual cumulative precipitation (1978 ~ 2008) is 462.7 mm with slightly more than 50% falls as rain in sumer (mean July = 63.3 mm; mean August = 74.1 mm; mean September = 71.5 mm).

Table 1. Maximum, Minimum and Mean of Annual Cumulative Precipitation (P_{cum}) and Annual Average Air Temperature (T_{avg}) over a 31-year Period at the Churchill Airport (1978-2008, Environment Canada)

Period	Maximum		Minimum		Mean	
	P_{cum} (mm)	T_{avg} (^{o}C)	P_{cum} (mm)	T_{avg} (^{o}C)	P_{cum} (mm)	T_{avg} (^{o}C)
1978-1982	603.0	-4.9	340.9	-8.5	470.2	-7.2
1983-1987	618.4	-5.4	412.7	-7.4	481.6	-6.7
1988-1992	522.8	-7.1	357.2	-8.2	440.8	-7.6
1993-1997	507.8	-6.6	290.5	-7.4	396.1	-7.0
1998-2002	596.6	-3.9	439.1	-7.4	504.0	-5.2
2003-2008	692.3	-3.6	351.6	-8.7	483.0	-5.8
1978-2008	692.3	-3.6	290.5	-8.7	462.7	-6.6

3. Modelling of the Deer River Watershed

3.1. The SLURP Hydrological Model

The SLURP model, version 11 (Kite, 1997) is a semi-distributed, conceptually continuous hydrological model which fits between the traditional lumped models and fully-distributed models. This daily time-step model was originally developed for meso- and macroscale basins with intermediate complexity, which incorporates necessary physical processes without compromising the simplicity of calculation. The model requires that the watershed must be divided into multiple aggregated simulation areas (ASAs) by TOpographic PArameteriZation (TOPAZ). After a preliminary removal of depression and flat areas from the input DEM, the D8 flow algorithm (O'Callaghan and Mark, 1984) is used to compare the elevation of each grid against the elevations of its eight neighbour grids and define a single flow direction to its steepest descent neighbour grid. Those raster grids that have upstream drainage area larger than the predefined critical source area (CSA) will be determined as the channel network. A modification of the channel network is further applied to eliminate any links shorter than the Preset Minimum Source Channel Length (MSCL) values. Eventually the channel links are ordered using the Strahler stream-ordering system (Strahler, 1957) and their contributing areas are indentified to generate the ASAs. Each ASA is further divided into areas with different types of land cover based on vegetation, soil, and physiographical conditions.

The vertical water balance module is applied to each type of land cover within an ASA using four tanks representing the canopy interception, snowpack, aerated soil storage, and groundwater. Precipitation intercepted by canopy is mainly affected by the amount of precipitation and the leaf area index (LAI) of the canopy. Excess precipitation that passes through canopy is counted as snow or surface storage on the basis of the melting temperature. Snow melt is calculated using either the degree-day method (Anderson, 1973) or the simplified energy budget method (Kustas et al., 1994). Snowmelt rate is interpolated (parabolically) between the snowmelt rates in January and July. Infiltration from snowmelt or rainfall to the subsurface storage tanks is governed by the Philip expression (Philip, 1954). Three methods from Morton (1983), Spittlehouse (1989), and Granger (1995) are available for calculating evapotranspiration, depending on data availability. Surface runoff is generated when water remaining in the aerated soil layers after infiltration exceeds the depression storage capacity. Subsurface flow (interflow and groundwater flow) is simulated at a rate depending on the water content of the subsurface storage tanks as well as the water transfer coefficient. Within each ASA, generated surface and subsurface flow is routed from each land cover to the nearest channel based on Manning's equation with different coefficients of roughness, hydraulic radiuses, and elevation changes. Runoff routing between ASAs is sequentially carried out by using either the hydrological storage techniques or the Muskingum-Cunge channel routing method. Travel time along channel to the final outlet is computed based on average distance to the outlet, surface slope, and flow velocity over each land cover. The accumulated flow from the outlet of an ASA is routed to the downstream ASA and finally to the outlet of the watershed.

3.2. Input Variables and Model Initialization

A 3-arc-second digital elevation model (DEM) of the Deer River watershed was obtained from the National Map Seamless Server of the U.S. Geological Survey (USGS, 2008). The DEM was processed by TOPAZ which has strong capability of automated digital landscape analysis to help delineate the sub-basins and drainage network. Meteorological records (1978 ~ 1997) at the Churchill-A Climate station (ID 5060600) were provided by Environment Canada. Streamflow data (1978 ~ 1997) were obtained from Water Survey Canada at the D. River N. Belcher station (ID: 06FD002, Figure 1) at which modelling results were compared with historical records. Land cover datasets were obtained from the Systeme Probatoire d'Observation dela Tarre (SPOT) earth observation satellite system (SPOT Vegetation Program, 2008) and reclassified into six land cover classes, including water, impervious, marsh, shrub, coniferous trees, and deciduous trees. The helicopter recons were also carried out on June 20 and Oct. 3, 2007 to collect information about vegetation coverage, topographic and hydrological conditions across the watershed, particularly in the upper reach of the Deer River.

The DEM and land cover information of the Chesnaye sub-basin were extracted from those of the Deer River watershed. Meteorological data were obtained from the automated weather station at Rail Spur. Data were scanned by a Campbell Scientic data logger (model CR1000) and stored as hourly averages. Frost table and surface soil moisture at multiple transects (2, 4, 6 and 8 m away from the bank) of each stream gauging station were measured using steel pole and the SM200 soil moisture sensor, respectively. Streamflow was monitored at the gauging stations deployed within the Chesnaye sub-basin using HOBO® water pressure transducer and Sontek® ADV Flowtracker.

According to data availability, snowpack was depleted using the degree-day method when air temperature was above the critical rain/snow division temperature; evapotranspiration was estimated using the Morton CRAE model (Bashir et al., 2009); and channel routing between ASAs was conducted using the storage routing method.

3.3. Sensitivity Analysis

To understand which model parameters contribute most to the output, sensitivity analysis was conducted by adjusting all the 22 model parameters (one-factor-at-a-time) and evaluating their impacts on modelling performance (Table 2). The base values of these parameters were determined in reference to field investigation of this research, the SLURP manual and other researchers' work in the HBL where the study area is located (Kite, 1997; Su et al., 2000; Metcalfe and Buttle, 2001; Kite, 2002; Thorne, 2004; Woo and Thorne, 2006). Based on the assumption of independent interrelationship, each parameter was individually adjusted by ±5%, ±15%, and ±30% while keeping others at their initial values (adjustment of each parameter was made to all land covers at one time), and evaluated by the fluctuation of the 10-year (1978 ~ 1987) logarithmic Nash and Sutcliffe efficiency (NSE) to derive a limited number of most influential ones. As comparing to the traditional NSE criterion, the logarithmic NSE (NSE_{ln}) uses logarithmic values of observed and simulated flows in order to enhance the influence of low flows without compromising the significance of peak flows. The 10-year logarithmic NSE was calculated by the following equation:

$$NSE_{ln} = 1 - \frac{\sum(\ln Q_0 - \ln Q_m)^2}{\sum(\ln Q_0 - \ln Q_{average})^2} \tag{1}$$

where Q_0 is the daily observed flow in 10 years (m^3/s); Q_m is the daily modeled flow in 10 years (m^3/s); and $Q_{average}$ is the 10-year mean observed flow (m^3/s). NSE_{ln} value is less than or equal to 1. The closer it is to 1, the better performance of the model.

Initial contents of snow store and slow store, maximum infiltration rate, retention constants for fast store and slow store, maximum capacity for fast store and slow store, precipitation factor, rain/snow division temperature, and snowmelt rates in January and July were determined as the most influential parameters (Table 2). These parameters will be included in model

Table 2. Sensitivity Analysis of Model Parameters

Adjustment of parameters (one at a time)		↓30%	↓15%	↓5%	↑5%	↑15%	↑30%
Parameters	Unit	Variation of logarithmic NSE (%)					
Initial contents of snow store	(mm)	+1.7	+0.8	+0.4	-0.3	-0.8	-1.4
Initial contents of slow store	(%)	+3.6	+1.7	+0.6	-0.5	-1.6	-3.0
Maximum infiltration rate	(mm/day)	+1.5	+0.6	+0.2	-0.1	-0.4	-0.7
Manning's roughness coefficient	(dimensionless)	0	0	0	0	0	0
Retention constant for fast store	(day)	+6.3	+3.2	+1.1	-0.9	-3.2	-6.4
Maximum capacity for fast store	(mm)	-3.2	-1.6	-0.4	+0.5	+1.5	+2.8
Retention constant for slow store	(day)	+1.9	+0.6	+0.2	+0.1	-0.2	-0.1
Maximum capacity for slow store	(mm)	+21	+17.4	+14.3	+12.8	+11.4	+9.0
Precipitation factor	(dimensionless)	+14.5	+19.9	+13.6	+7.1	+3.3	-10.0
Rain/snow division temperature	(^{o}C)	+12.2	+13.3	+13.9	+14.5	+15.1	+15.7
Canopy interception A	(dimensionless)	+2.4	+0.7	+0.2	0	0	0
Canopy interception B	(dimensionless)	+3.1	+2.1	+0.3	0	0	0
Land cover albedo	(dimensionless)	0	0	0	0	0	0
LAI in January	(dimensionless)	-0.3	-0.1	-0.1	+0.1	+0.1	+0.3
LAI in July	(dimensionless)	-0.1	-0.1	-0.1	0	0	-0.1
Maximum canopy capacity	(dimensionless)	-0.2	-0.1	-0.1	+0.1	+0.1	+0.1
Soil heat flux amplitude	(dimensionless)	0	0	0	0	0	0
Snowmelt rate in January	(mm/^{o}C/day)	-7.3	-3.5	+0.6	-0.4	-1.9	-8.7
Snowmelt rate in July	(mm/^{o}C/day)	-5.8	-2.7	-0.8	+0.4	+0.5	-1.1
Maximum albedo of snow	(dimensionless)	0	0	0	0	0	0
Minimum albedo of snow	(dimensionless)	0	0	0	0	0	0
Temperature lapse rate	(^{o}C/100 m)	0	0	0	0	0	0

calibration except for the precipitation factor, which is used to compensate the precipitation gauge and was set to 1.0 for no compensation. Other insignificant parameters were not addressed in the calibration and their base values were used as the final values for modelling validation.

3.4. Modelling Validation

The SLURP model can only optimize the first 10 parameters shown in Table 3 through its built-in automatic calibration module. Therefore, calibration was conducted for the first 10-year period (1978 ~ 1987) involving both the built-in module and manual adjustment which focused on the snowmelt rates in January and July. The manual adjustment was based on the results from sensitivity analysis and field investigation. The initial values of parameters that require optimization were adopted in reference to those based values used in the sensitivity analysis. The based values were obtained from field investigation of this research, the SLURP manual and other researchers' work in the HBL where the study area is located. Both of the NSE_{ln} criterion and the deviation of runoff volumes (DV) were used as statistical measures of the goodness of fit of the modelling results at the D. River N. Belcher station. Table 3 summarizes the final calibrated parameters for the Deer River watershed from 1978 to 1987. Snowmelt rates in January were optimized as 0 mm/^{o}C/day to minimize the magnitude of premature snowmelt runoff. Snowmelt rates in July were manually set to 4, 3 and 2 mm/^{o}C/day for water/impervious/marsh, shrub, and coniferous/deciduous areas, respectively (Metcalfe and Buttle, 2001).

Calibration of the SLURP model to the Deer River watershed in 1980 resulted in an NSE_{ln} of 71% and an average DV of -33% (Table 4). The year 1980 was a median year from a meteorological perspective and therefore was selected as an example for further discussion. Of the 29% deficiency in NSE_{ln}, the majority was attributed to the simulation during spring snowmelt and summer, whereas the minority was contributed during fall and winter (Figure 2). The simulated snowmelt peak runoff occurred 2 days earlier and it was 24% lower than the observed spring peak. Streamflow during summer and fall months was drastically underestimated by the SLURP model, with the cumulative flow 64% less than the observed discharge. An-

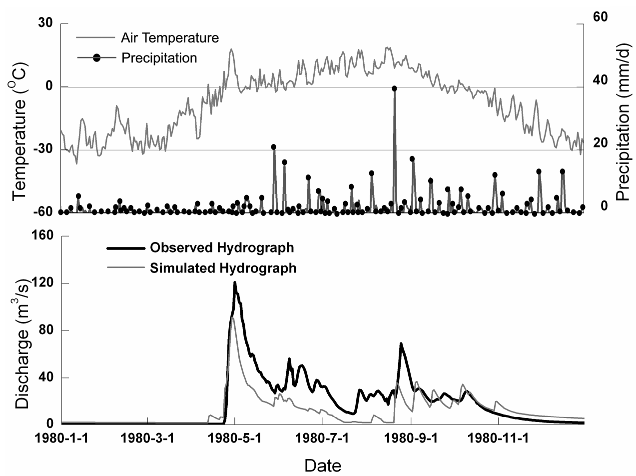

Figure 2. Simulated and observed daily hydrographs for the Deer River watershed in 1980.

Table 3. Final Values of the Model Parameters for Each Land Cover Type in the Deer River Watershed (Italicized parameters were automatically or manually optimized during model calibration)

Parameters	Unit	Water	Impervious	Marsh	Shrub	Coniferous
Initial contents of snow store	(mm)	1	1	1	1	1
Initial contents of slow store	(%)	9.775	8.625	4.238	6.839	5.895
Maximum infiltration rate	(mm/day)	100.9	142.4	106.9	147.7	111.9
Manning roughness n	(dimensionless)	0.02	0.08	0.01	0.07	0.02
Retention constant for fast store	(day)	36.97	52.62	5.447	7.480	62.89
Maximum capacity for fast store	(mm)	95.35	133.8	531.2	583.6	373.7
Retention constant for slow store	(day)	130.7	171.0	686.1	745.5	713.0
Maximum capacity for slow store	(mm)	338.7	260.6	361.6	102.9	62.19
Precipitation factor	(dimensionless)	1	1	1	1	1
Rain/snow division temperature	(^{o}C)	-0.03	-0.56	-0.99	-0.93	-0.61
Canopy interception A	(dimensionless)	0	0.5	1	1	1
Canopy interception B	(dimensionless)	1	1	1	1	1
Land cover albedo	(dimensionless)	0	0.15	0.15	0.14	0.13
LAI in January	(dimensionless)	0	0	2	0.5	5
LAI in July	(dimensionless)	0	2	2	4.5	5
Maximum canopy capacity	(dimensionless)	0	2.8	3.8	6.2	5.6
Soil heat flux amplitude	(dimensionless)	0.15	0.15	0.15	0.15	0.15
Snowmelt rate in January	(mm/^{o}C/day)	0	0	0	0	0
Snowmelt rate in July	(mm/^{o}C/day)	4	4	4	3	2
Maximum albedo of snow	(dimensionless)	0.78	0.78	0.78	0.7	0.7
Minimum albedo of snow	(dimensionless)	0.37	0.37	0.37	0.37	0.37
Temperature lapse rate	(^{o}C/100 m)	0.75	0.75	0.75	0.75	0.75

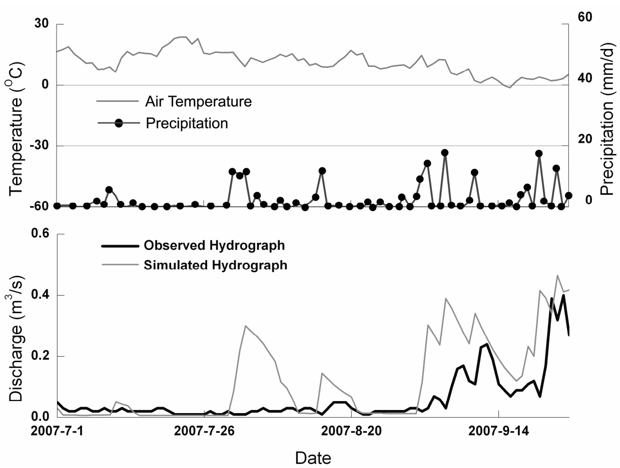

Figure 3. Simulated and observed daily hydrographs by the SLURP model for Station 6 in 2007.

nual DV represents the difference between standard deviations of both simulated and observed runoff, indicating whether the SLURP model overestimates or underestimates the runoff. Although the year of 1980 had a negative DV of -33%, most of the annual DV values during the calibration years were positive, indicating that runoff volumes for the majority of the calibration years were overestimated. This overestimation may be attributed to the combined effects from not considering the existence of permafrost, and underestimating the water storage capacity of organic soil layers.

Modelling verification was performed for another period of 10 years (1988 ~ 1997). The simulated hydrographs for the verification years demonstrated that snowmelt peak runoff occurred 2 ~ 8 days later and the peak runoffs were 34% lower in average as compared to the observed records, indicating the simulated snowpack was not depleting as quickly as it was in nature. This deviation may be due to the inherent date-dependent snowmelt rates used in the degree-day method. The degree-day method generates runoff when air temperature exceeds the predefined rain/snow division temperature. To reduce the sudden runoff response without compromising modelling accuracy, snowmelt rates were manually adjusted as low as possible which in turn delays and extends the melting period. On the other hand, streamflow was overestimated during summer and fall when rainfall dominates the wetland water recharge and permafrost descends. Most annual DV values of the verification years were also positive which supports these conclusions as well.

3.5. The Chesnaye Sub-basin

To further test the model's applicability in micro-scale basin, validation was conducted for the Chesnaye sub-basin from 2006 to 2008. Values of model parameters were adopted from the calibration results for the Deer River watershed to maintain the hydrologic consistency (Table 3). In spite of reasonable modelling performance, the results indicated that runoff generated from most rainfall events were overestimated in summer. Canopy interception, depression storage, organic-mineral soil layer mixture, permafrost and evapotranspiration may explain the fact that water was mainly stored in ponds and soil layers rather than discharged as stream runoff. On the other hand, runoff from rainfall events was precisely estimated in late fall,

Table 4. Summary of Annual Modelling Outputs at the D. River N. Belcher Station for the Deer River Watershed (1978 -1997)

Year	CP* (mm)	CET (mm)	MT (°C)	OMS (m³/s)	MS (m³/s)	NSE_{ln} (%)	DV (%)
1978	532.8	218.7	-7.76	12.7	14.6	71	16
1979	341.9	194.6	-7.96	13.9	9.9	51	-29
1980	484.0	242.3	-6.93	18.4	12.4	71	-33
1981	395.4	203.1	-4.89	14.4	17.9	61	24
1982	605.8	239.6	-8.43	15.8	24.6	41	56
1983	621.0	248.6	-7.06	27.9	35.7	68	28
1984	413.9	195.0	-6.50	14.3	17.8	43	25
1985	448.5	184.4	-7.11	11.1	17.0	38	53
1986	500.3	241.0	-7.33	22.8	25.6	57	12
1987	432.7	227.1	-5.42	11.6	17.1	42	47
1988	441.3	180.2	-7.38	9.07	6.68	72	-26
1989	358.5	172.6	-8.02	17.6	15.8	70	-10
1990	485.3	187.4	-7.41	12.1	16.4	22	36
1991	524.9	202.0	-7.09	23.8	28.6	62	20
1992	402.0	169.3	-8.21	12.8	23.1	15	80
1993	291.5	177.6	-7.11	9.0	10.4	13	16
1994	345.5	147.4	-6.62	7.2	14.0	21	96
1995	416.4	224.1	-6.90	19.4	17.2	59	-12
1996	424.1	200.7	-7.46	7.0	12.7	25	82
1997	509.7	185.4	-6.55	27.2	22.1	68	-19
Average	448.8	202.1	-7.11	15.4	18.0	49	±36

**CP* is cumulative precipitation; *CET* is cumulative evapotranspiration; *MT* is mean temperature; *MOS* is mean observed streamflow; *MSS* is mean simulated streamflow.

which indicates that soil layers were saturated after the summer drainage while the storage capacity was declining (Figure 3). These results agreed with the conclusion from modelling the Deer River watershed (macro-scale) that the SLURP model overestimated the runoff from summer rainfall events due to the lack of consideration of the dynamics of permafrost and ponds (Table 5).

4. Results and Discussion

The results from modelling the Deer River watershed indicated that snowmelt in the spring season (April-June) produced approximately half of the annual water recharge. This finding revealed that, in subarctic wetlands, snow accumulation is the major source of surface water. Peaks of the simulated spring runoffs, as shown in Figure 4, were 34% lower than the observed ones in average during the modelling period. This could be attributed to the combined effects of the following factors. Snow sublimation was considered as a part of snowmelt rather than an independent process in the SLURP model. However, it could remove great amount of snow content from producing spring runoff. Permafrost represents an over-winter surface storage of groundwater. When temperature increases and snow starts to melt, water is also released as stream runoff from the thaw of ice contents within the soil layers. Moreover, permafrost can prevent water from penetrating into deep soil layers and therefore amplify the actual spring runoff.

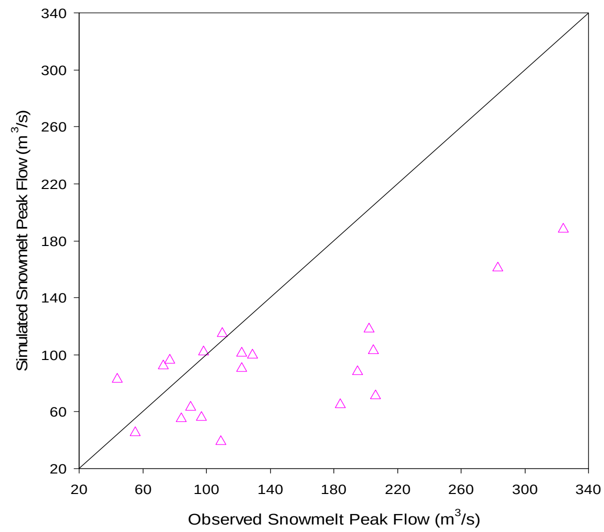

Figure 4. Observed and simulated snowmelt peak flows at the D. River N. Belcher station (1978-1997).

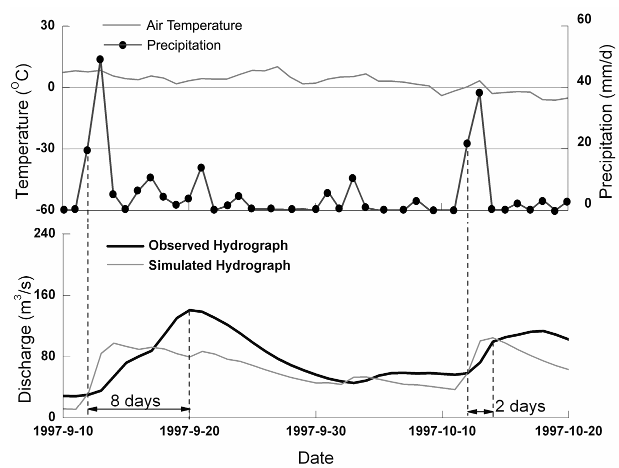

Figure 5. Response of daily discharge to the precipitation by the SLURP model in 1997.

Both simulated and observed streamflow showed that most light and moderate rainfall events in summer (July-September) were not able to generate notable runoff. This phenomenon may be due to various reasons, including canopy interception, depression storage, porous soil layers, impermeable permafrost and intensive evapotranspiration. The dominant vegetation species in the Deer River watershed are tundra, shrub, and coniferous forest which have considerable interception capacities. Depression storage is referred to the numerous ponds which store great amount of water and become connected in wet seasons. Water levels of these ponds fluctuate significantly with climatic conditions and wetland water budget in various seasons. The Deer River watershed has highly porous soil which allows water to infiltrate into deep soil layers and finally to be discharged through evapotranspiration and groundwater flow. Descending frost table in the summer releases more porous soil layers and therefore enlarges the total water storage capacity. The average annual cumulative evapotranspiration is 45% of

Table 5. Summary of Modelling Outputs at Stations 5, 6, 7 and 10 for the Chesnaye Sub-basin from 2006 to 2008

Year	S#*	SP	CP (mm)	CET (mm)	MT (°C)	MOS (m^3/s)	MSS (m^3/s)
06	5	Jul 1- Aug 20	135.6	79.7	13.9	0.14	0.17
	7	Jul 1- Aug 20	135.6	79.7	13.9	0.11	0.50
07	5	Jul 1- Sep 26	189.0	73.7	11.2	0.56	0.20
	6	Jul 1- Sep 26	189.0	73.7	11.2	0.06	0.12
	7	Jul 28- Sep 26	172.2	56.1	9.2	0.30	0.80
	10	Jul 1- Aug 21	75.8	52.4	14.5	0.13	0.04
08	5	Jul 13- Aug 8	44.9	22.3	15.3	0.55	0.09
	6	Jul 13- Sep 21	108.3	54.7	12.5	0.17	0.07
	10	Jul 13- Sep 17	97.1	53.1	13.1	0.45	0.04

*S# is station number; SP is simulation period; *CP* is cumulative precipitation; *CET* is cumulative evapotranspiration; *MT* is mean temperature; *MOS* is mean observed streamflow; *MSS* is mean simulated streamflow.

the average annual cumulative precipitation, indicating that evapotranspiration also dominates the water cycle of the Deer River watershed especially in summer (Table 4). Evapotranspiration also tends to be intensified due to higher air temperature and longer daylight period in summer and therefore further reduces surface runoff. These combined factors resulted in the fact that only heavy or continuous rainfall events were able to generate countable runoff. Rainfall events that occurred in the fall generate much more runoff due to relatively low temperature and less net radiation.

Modelling results also showed a lag of 2 ~ 8 days between the peaks of rainfall and runoff in both summer and fall. As shown in Figure 5, a short-duration (30 hours) and high-intensity (59 mm in total) rainfall event occurred on October 12 and 13, 1997. Both simulated and observed streamflow peaks appeared on October 14. Another series of high-intensity and moderate (2 ~ 49 mm) rainfall events occurred from September 12 to 20 with an 8-day delay of observing the peak flow. The large buffe- ring capacity of wetland water storage plays a key role during the runoff concentration. A high-intensity rainfall event brings plenty of water to the wetland in a short period. After the surface soil layer is saturated, excess water generates flashy runoff, resulting in more rapid runoff response. On the other hand, if rainfall events are concentrated but moderate, infiltration dominates water distribution and allows water to penetrate into deep soil layers from which it could be gradually routed to the streams as groundwater flow. Therefore, the runoff response is prolonged and much gentler. Particularly, numerous ponds stretching over the watershed behave as buffers and further extend the concentration time.

Although the modelling results reasonably matched the observed data, some limitations should be noticed. Using meteorological data from the town of Churchill, which is 70 km north to the watershed, could influence the hydrological processes and compromise the modelling accuracy. The majority of the watershed is plain wetland with slightly varying elevation and the resolution of the DEM obtained from the USGS is 90 m. Although relatively low resolution DEM would affect the simulation accuracy of precipitation-runoff responses, many previous studies have reported acceptable results using 90 m (or even lower) resolution DEM for small to medium size watersheds (Van der Linden and Woo, 2003; St Laurent and Valeo, 2007; Armstrong and Martz, 2008). Resolution of the NDVI data obtained from the SPOT vegetation program is 1 km. To match the resolution of the DEM, each NDVI value was uniformly distributed to multiple DEM grids. This conversion may sacrifice the accuracy of land cover classification. Moreover, there are a number of small ponds, which are not detected in the SPOT image, may influence the modelling outputs. Calibration was implemented through both the built-in optimization module and manual adjustments. However, the values of some parameters (e.g., snowmelt rate, infiltration rate) may not precisely reflect the actual conditions in the study area. This problem could be mitigated if these parameters were obtained from the field measurement.

5. Conclusions

To better understand the hydrological characteristics of subarctic wetlands, a semi-distributed, conceptually continuous hydrological model - the SLURP model, was applied to simulate the hydrology of the Deer River watershed in northern Manitoba. Sensitivity analysis showed that maximum capacity for slow store, rain/snow division temperature, and snowmelt rates were the most influential parameters. Calibration and validation of the model produced an average NSE_{ln} of 49% and DV of ±36% as well as a number of interesting findings. The modelling results indicated that snowmelt in the spring season is the major water replenishment and constitutes approximately 50% of the annual runoff in the Deer River watershed. Simulated snowmelt peak flows were 34% less than the historical records in average which could be attributed to the effects of permafrost. The shallow permafrost could act as an effective barrier for infiltration and therefore amplify the spring runoff. Most of the light and moderate summer rainfall events were not able to generate notable runoff due to canopy interception, depression storage, porous soil layers, declining frost table, and intensive evapotranspiration. The average annual cumulative evapotranspiration was 45% of the average annual cumulative precipitation, indicating the significance of evapotranspiration to the Deer River watershed especially in the summer months. Contrastingly, rainfall events that occurred in fall produced more surface runoff because the decreasing temperature and net radiation alleviated the intensity of evapotranspiration. The modelling results also showned a lag of 2 ~ 8 days between peaks of rainfall and runoff during summer and fall, demonstrating a considerable water storage capacity of the organic soil layers and buffering effect of the ponds. The simulation results obtained in the Chesnaye sub-basin further indicated that the SLURP model overestimated rainfall events due to the lack of considering permafrost and seasonal ponds as well as underestimating evapotranspiration. Modifications to the subroutines of snowmelt, permafrost, and depression storage would be able to improve the model's performance in the subarctic wetland domain. Despite the crude resolution and uncertainties of the inputs, this study advanced the knowledge about the climatic, geographical and hydrological characteristics of the Canadian subarctic wetlands through the simulation by the SLURP model.

Acknowledgments. Special thanks go to Manitoba Hydro, ArcticNet, NFSC (50709010), the Major State Basic Research Development Program of MOST (2006CB403307), NSERC, and Churchill Northern Studies Center (CNSC) for funding this work. The advice and help from Dr. Kenneth Snelgrove at Memorial University of Newfoundland and Dr. Tim Papakyriakou at the University of Manitoba were also highly appreciated.

References

Abbott, M.B., Bathurst, J.C., Cunge, J.A., O'Connell, P.E., and Rasmussen, J. (1986). An Introduction to the European System: Systeme Hydrologique Europeen (SHE). *J. Hydrol.,* 87, 61-77. doi: 10.1016/0022-1694(86)90115-0

ACIASC. (2004). *Impacts of a Warming Arctic: Arctic Climate Impact Assessment.* Cambridge University Press, Cambridge, UK.

Armstrong, R.N., and Martz, L.W. (2008). Effects of reduced land cover detail on hydrological model response. *Hydrol. Process.*, 22, 2395-2409. doi:10.1002/hyp.6830

Anderson, E.A. (1973). National Weather Service River Forecast System-Snow Accumulation and Ablation Model. *A Technical Report NWS 19, National Oceanographic and Atmospheric Administration (NOAA)*, USA.

Bashir, M.A., Tanakamaru, H., and Tada, A. (2009). Spatial and Temporal Analysis of Evapotranspiration Using Satellite Remote Sensing Data: A Case Study in the Gezira Scheme, Sudan. *J. Env. Inform.*, 13(2), 86-92. doi:10.3808/jei.200900143

Bergström, S. (1992). The HBV model - its structure and applications. *SMHI*, Norrköping, Reports RH, No. 4.

Beven, K.J., Kirkby, M.J., Schofield, N., and Tagg, A.F. (1984). Testing a physically-based flood forecasting model (TOPMODEL) for three U.K. Catchments. *J. Hydrol.*, 69, 119-143.

Bicknell, B.R., Imhoff, J.C., Kittle, J.L., Donigian, A.S., and Johanson, R.C. (1997). Hydrological Simulation Program-Fortran, User's manual for version 11: U.S. Environmental Protection Agency. *National Exposure Research Laboratory*, Athens, Ga., EPA/600/R-97/080, 755 p.

Boswell, J.S., and Olyphant, G.A. (2007). Modelling the hydrologic response of groundwater dominated wetlands to transient boundary conditions: Implications for wetland restoration. *J. Hydrol.*, 332, 467-476. doi:10.1016/j.jhydrol.2006.08.004

Chen, B. (2007) Climate change and pesticide losses in watershed systems: a simulation modeling study. *J. Env. Inform.,* 10(2), 55-67. doi:10.3808/jei.200700100

Chen, B. (2008) Health-impact risk assessment for river pollution through a modified contaminant transport model. *Water Air Soil Poll.,* 200(1-4), 323-339. doi:10.1007/s11270-008-9915-5

Chen, B., Li, Y.F., Huang, G.H., Struger, J., Zhang, B.Y., and Wu, S.M. (2003) Estimation of atrazine loss through the surface runoff from agricultural land in Auglaize-Blanchard Watershed, the United States. *Water Qual. Res. J. Can.*, 38(4), 585-606.

Chen, B., Li, Y.F., Huang G.H., Huang, Y.F., and Li, Y.R. (2004) PELM: an integrated pesticide losses model for simulating pesticide pollution in a watershed system. *J. Environ. Sci. Health - Part B*, B39(4): 613-626. doi:10.1081/PFC-200026853

Granger, R.J. (1995). A feedback approach for the estimation of evapotranspiration using remotely sensed data. In: Kite, G. W., Pietroniro, A. and Pultz, T. (Eds), Applications of Remote Sensing in Hydrology, *Proceedings of the Second International Workshop*, 18-20 October, 1994. NHRI, Saskatoon, 211-222.

Jing, L. (2009). Hydrological Monitoring and Modelling of the Deer River watershed with SLURP and WATFLOOD. *M.Eng. thesis*. Faculty of Engineering and Applied Science, Memorial University of Newfoundland, St. John's, NL, Canada.

Jing, L., Chen, B., Snelgrove, K., and Papakyriakou, T. (2009). Hydrological Monitoring and Characterization of a Sub-Arctic Wetland in the Hudson Bay Lowlands. *Proceedings of the Canadian Society of Civil Engineering 2009 Annual Conference*, St. John's, Canada, GC-135.

Kite, G.W. (1997). *Manual for the SLURP Hydrological Model V.11.* National Hydrology Research Institute, Saskatoon, Canada.

Kite, G.W. (2002). *Manual for the SLURP hydrological model, v12.2.* International water management institute, Colombo, Sri Lanka.

Kouwen, N., Soulis, E.D., Pietroniro, A., Donald, J., and Harrington, R.A. (1993). Grouped response units for distributed hydrologic modeling. *J. Water. Res. Pl-ASCE.,* 119, 289-305.

Kustas, W.P., Rango, A., and Uijlenhoet, R. (1994). A simple energy budget algorithm for the snowmelt runoff model. *Water Resour. Res.*, 30(5), 1515-1527. doi:10.1029/94WR00152

Leenders, E.E., and Woo, M.K. (2002). Modelling a two-layer flow system at the subarctic, subalpine treeline during snowmelt. *Water Resour. Res.*, 38(10), 1202. doi:10.1029/2001WR000375

Li, P., Chen, B., and Husain, T. (2010). IRFAM: an Integrated Rule-Based Fuzzy Adaptive Resonance Theory Mapping System for Watershed Modeling. *ASCE J. Hydrol. Eng.*, American Society of Civil Engineers. doi:10.1061/(ASCE)HE.1943-5584.0000285

Mansell, R.S., Bloom, S.A., and Sun, G. (2000). A Model for Wetland Hydrology: Description and Validation. *Soil Sci.*, 165(5), 384-397. doi:10.1097/00010694-200005000-00002

Metcalfe, R.A., and Buttle, J.M. (2001). Soil partitioning and surface store controls on spring runoff from a boreal forest peatland basin in north-central Manitoba, Canada. *Hydrol. Process.*, 15, 2305-23 24. doi:10.1002/hyp.262

Morton, F.I. (1983). Operational estimates of areal evapotranspiration and their significance to the science and practice of hydrology. *J. Hydrol.*, 66, 77-100. doi:10.1016/0022-1694(83)90178-6

Mills, G.F., Veldhuis, H., Forrester, D.B., and Schmidt, R. (1976). A guide to bio-physical land classification in Manitoba. *Northern Resource Information Program*, Department of Renewable Resources and Transportation Services and Canada-Manitoba Soil Survey.

O'Callaghan, J.F., and Mark, D.M. (1984). The extraction of drainage networks from digital elevation data. *Comput. Vision, Graph. Image Process.*, 28, 323-344. doi:10.1016/S0734-189X(84)80011-0

Payette, S., Fortin, M.J., and Gamache, I. (2001). The Subarctic Forest-Tundra: The Structure of a Biome in a Changing Climate. *Bioscience*, 51(9), 709-718. doi:10.1641/0006-3568(2001)051[07 09:TSFTTS]2.0.CO;2

Philip, J.R. (1954). An infiltration equation with physical significance. *Soil Sci.*, 77(1), 153-157. doi:10.1097/00010694-195402000-00009

Price, J.S., and Waddington, J.M. (2000). Advance in Canadian wetland hydrology and biochemistry. *Hydrol. Process.*, 14, 1579-1589. doi:10.1002/1099-1085(20000630)14:9<1579::AI D-HYP76 >3.0.CO;2-#

Price, J.S., Branfireun, B.A., Waddington, J.M., and Devito, K.J. (2005). Advances in Canadian wetland hydrology, 1999-2003. *Hydrol. Process.*, 19, 201-214. doi:10.1002/hyp.5774

Quick, M.C., and Pipes, A. (1977). U.B.C. Watershed Model. *Hydrol. Sci. - Bulletin*, XXII(1), 153-161. doi:10.1080/02626667709491701

Quinton, W.L., and Roulet, N.T. (1998). Spring and Summer Runoff Hydrology of a Subarctic Patterned Wetland. *Arctic Alpine Res.*, 30(3), 285-294. doi:10.2307/1551976

Reeve, A.S., Siegel, D.I., and Glaser, P.H. (2000). Simulating vertical flow in large peatlands. *J. Hydrol.*, 227, 207-217. doi:10.1016/ S00 22-1694(99)00183-3

Richard, C., and Gratton, D.J. (2001). The importance of the air temperature variable for the snowmelt runoff modelling using the SRM. *Hydrol. Process.*, 15, 3357-3370. doi:10.1002/hyp.1031

Rouse, W.R., Douglas, M.S.V., Hecky, R., Hershey, A.E., Kling, G.W., Lesack, L., Marsh, P., Mcdonald, M., Nicholson, B.J., Roulet, N.T., and Smol, J.P. (1997). Effects of climate change on the fresh waters of arctic and subarctic northern America. *Hydrol. Process.*, 11, 873-902.

Spittlehouse, D.L. (1989). Estimating evapotranspiration from land surfaces in British Columbia. *In: Estimation of Areal Evapotranspiration, IAHS Publication*, 177, 245-253. doi:10.1002/(SICI)1099-1085 (19970630)11:8<873::AID-HYP510>3.0.CO;2-6

SPOT Vegetation Program. The image processing and archiving centre, Mol, Belgium. http://www.vgt.vito.be (accessed June 23, 2008).

St Laurent, M.E., and Valeo, C. (2007). Large-scale distributed watershed modelling for reservoir operations in cold boreal regions. *Can. J. Civ. Eng.*, 34, 525-538. doi:10.1139/L06-142

Strahler, A.N. (1957). Quantitative analysis of watershed geomorphology. *Trans. Am. Geophys. Union*, 38(6), 913-920.

Ström, L., and Christensen, T.R. (2007). Below ground carbon turnover and greenhouse gas exchanges in a sub-arctic wetland. *Soil boil. Bioch.*, 39, 1689-1698.

Su, M., Stolte, W.J., and Van der Kamp, G. (2000). Modelling Canadian prairie wetland hydrology using a semi-distributed streamflow model. *Hydrol. Process.*, 14, 2405-2422. doi:10.1002/1099-1085(20001015)14:14<2405::AID-HYP92>3.0.CO;2-B

Thorne, R. (2004). Simulating streamflow of a large mountainous catchment using different sets of climatic data. *M.Sc. thesis*, McMaster University, Hamilton, Ontatio, Canada.

USGS. The National Map Seamless Server. http://seamless.usgs.gov/index.php (accessed June 23, 2008).

Van der Linden, S., and Woo, M.K. (2003). Application of hydrological models with increasing complexity to subarctic catchments. *J. Hydrol.*, 270, 145-157. doi:10.1016/S0022-1694(02)00291-3

Wessel, D.A., and Rouse, W.R. (1994). Modelling Evaporation from Wetland Tundra. *Boundary Layer Meteorol.*, 68, 109-130. doi:10.1007/BF00712666

Woo, M.K., and Marsh, P. (2005). Snow, frozen soils and permafrost hydrology in Canada, 1999-2002. *Hydrol. Process.*, 19, 215-229. doi:10.1007/BF00712666

Woo, M.K., and Thorne, R. (2006). Snowmelt contribution to discharge from a large mountainous catchment in subarctic Canada. *Hydrol. Process.*, 20, 2129-2139. doi:10.1002/hyp.6205

Woo, M.K., and Young, K. (2006). High Arctic wetlands: Their occurrence, hydrological characteristics and sustainability. *J. Hydrol.*, 320, 432-450. doi:10.1016/j.jhydrol.2005.07.025

Zhang, Z., Kane, D.L., and Hinzman, D. (2000). Development and application of a spatially-distributed Arctic hydrological and thermal process model (ARHYTHM). *Hydrol. Process.*, 14, 1017-1044. doi:10.1002/(SICI)1099-1085(20000430)14:6<1017::AID-HYP982>3.0.CO;2-G