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Wetland Monitoring, Characterization and Modelling under Changing Climate in the Canadian Subarctic

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ABSTRACT. Subarctic wetlands that exist as bogs, fens, swamps, marshes and shallow water, comprise 3% of the Canadian landscape. They have been recognized as important ecotones between the arctic tundra and boreal forest. Recently, there has been growing research interest in the hydrological characteristics of arctic and subarctic wetland systems in the need for more efficiently conserving wetlands and assessing climate change related impacts. This research targets the Deer River watershed near Churchill, Manitoba, which represents a typical subarctic wetland system in the Hudson Bay Lowlands. An extensive field investigation was first conducted during the summer from 2006 to 2008 to facilitate in-depth understanding of the wetland hydrology. The results provided evidence to indicate a strong relationship between air temperature and evapotranspiration. Permafrost table, soil moisture and streamflow were monitored and analyzed to advance the acknowledgement of the climatic, geographical and hydrological characteristics of subarctic wetlands. To quantify the water cycle and further validate the findings from field investigation, a Canadian distributed hydrological model, WATFLOOD, was employed to simulate the hydrologic processes in the targeted watershed. The results demonstrated that snowmelt in the spring season (April-June) was the major source of water supplement of subarctic wetlands. Most light and moderate rainfall events in summer (July-September) generated relatively small amounts of runoff which can be related to canopy interception, depression storage, porous soil layers, impermeable permafrost and intensive evapotranspiration. A lag of 2-8 days between the peaks of rainfall and stream runoff was observed in both summer and fall. This study is expected to benefit wetland conservation and the assessment of climate change related impacts in the Canadian northern regions.

Keywords: subarctic wetland, field investigation, hydrological modeling, permafrost, snowmelt

1. Introduction

The subarctic region covers much of northern Canada and is often characterized by taiga forest vegetation with relatively mild winters (Petrescu et al., 2010). The taiga consists primarily of coniferous forest and is interspersed by lichen and wetland landscapes such as bog marsh and muskeg (Kitti et al., 2009). These wetlands, usually with rich vegetation growth, span almost 3% of the Canadian landscape and offer habitats for wild lives (Price and Waddington, 2000). Subarctic wetlands, therefore, have been recognized as important ecotones between the arctic tundra and boreal forest. The hydrological processes that create and maintain these wetlands as well as 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; Woo and Young, 2006; Ström and Christensen, 2007; Jing and Chen, 2011).

Winter and Woo (1990) stated that adequate water source was the primary factor of the existence of subarctic wetlands. Quinton and Roulet (1998) demonstrated the relationship between flux and water storage of a subarctic patterned wetland, conceptualizing the discharge response delay to precipitation which is attributed to large storage capacity of pools. Woo and Young (2006) also noted that reliable water supply which comprises of snowmelt water, localized ground water discharge, stream flow and inundation by lakes and sea during the thawed season plays a determinant role in wetland sustainability. Besides these, water flow within northern wetlands is highly sensitive to precipitation because of particular porous soil characteristic and shallow impermeable permafrost table. Woo and Marsh (2005), basing on reviewing the frozen soil and permafrost hydrology in Canada from 1999 to 2002, showed two distinctive flow mechanisms of subarctic wetland related to permafrost and permafrost table fluctuation. Hayashi et al. (2007) noted that subsurface flow is strongly dependent on permafrost table and developed a simulation method for hydrological models. Soil features of subarctic wetlands have been previously

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investigated by many studies and the preliminary results indicate that organic soil, which consists of acrotelm layer and catotelm layer, is underlain by mineral soil which has negligible capability of water infiltration (Carey and Woo, 1999; Woo and Marsh, 2005). Quinton and Marsh (1998) stated that hydraulic conductivity declines with depth because of increasing humidification of peat and moreover, Carey and Woo (1998) also found that discontinuity between organic and mineral layers leads to the explicit vertically hydraulic reduction. Carey and Woo (2000) studied on subarctic slopes and concluded that pipeflow is ephemeral when water table is within or above and diminishes during summertime when water table is drawn downward. Carey et al. (2007) estimated hydraulic and pore characteristics of organic soil in the Wolf Creek Basin, Yukon, educed that hydraulic conductivity and active layer porosity both decline with depth.

Yet, the studies on subarctic wetlands are always restrained by their limited occurrence, precautious existence and remoteness (Li et al., 2010). To help understand the hydrological processes, hydrological models have been widely used as simplified, conceptual representations of the hydrologic cycle (Filoso et al., 2004; Wu and Johnston, 2007; Schmalz et al., 2008; Hattermann et al., 2008; He et al., 2010; Gao et al., 2010; Ping et al., 2010). However, integrated research efforts on monitoring and modelling subarctic wetlands in the Hudson Bay lowlands (HBL) has been limited due to its physical accessibility and technical difficulties such as data availability. This paper examines the hydrological and vegetation characteristics of subarctic wetlands in the HBL, combining and discussing the findings from an extensive field investigation with the modelling results from the WATFLOOD hydrological model to provide a synopsis on the environment that influences the development and management of subarctic wetlands.

2. Study Area

This study was conducted at the Deer River watershed, a 5048 km^2 subarctic wetland in the northern part of the HBL, Manitoba, Canada (Figure 1). The study area is composed mostly of muskeg and peatlands, and dotted with seasonal ponds, lakes and streams. Elevation gradually descends from 232 m in the southwest to 16 m in the northeast (Jing et al., 2009). Spruce, birch alder, and balsam fir dominate the headwaters as well as the adjacent areas along river channels. Shrub and tundra prevails in the hummocky terrain within the mid- and downstream regions. The primary soils are brunisolic static cryosol, brunisols, brunisolic turbic cryosol, and organo cryosol with a peat depth of 1 to 3 m that in parts are underlain by continuous permafrost with an active depth of approximately 1 m by late August (Mills et al., 1976; Malmer and Wallén, 1996; Christensen et al., 2004). Subsurface water content reaches its maximum equilibrium following snowmelt when surface water fully recharges the soil layers. The Deer River watershed has been categorized as a marine subarctic climate. Mean annual precipitation at nearby Churchill is 462 mm with maximum in summer and mean annual temperature is -6.5 ^{o}C (1978 ~ 2007). Winters are long and cold with average temperature varying around -20 ^{o}C from November to April. Runoff is extremely low in winters and is mainly sustained by groundwater discharge. Spring returns in early or mid May when snowmelt starts and ceases by mid June which usually generates annual peak runoff. Soil water deficits are common in summer and fall due to the intensive evapotranspiration and lack of precipitation.

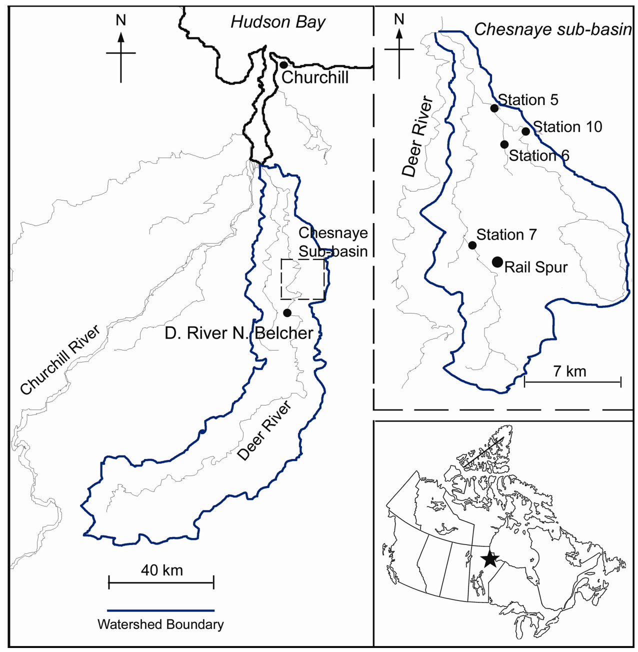

Figure 1. Location of the Deer River watershed and the Chesnaye sub-basin (with monitoring stations).

3. Methods

3.1. Field Investigation

Due to limited accessibility, a representative sub-basin in the lower reach of the Deer River, the Chesnaye sub-basin, was selected for an extensive field investigation during the summer time from 2006 to 2008 (Figure 1). 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 (Jing, 2009). Meteorological parameters such as air temperature, dew point temperature, cumulative precipitation, incident short wave radiation, relative humidity, and wind speed and direction, were obtained from the weather stations. Data were scanned by a Campbell Scientic data logger (model CR1000) and stored at hourly intervals. Evapotranspiration was estimated by the FAO-56 Penman-Monteith Equation (Allen et al., 1998). Streamflow was measured at each gauging station by HOBO® water pressure transducers and Sontek® ADV Flowtracker. Permafrost table and surface soil moisture were measured at multiple transects (2, 4, 6 and 8 m from both banks) of each station using steel pole and SM200 soil moisture sensor. The helicopter recons were also carried out on June 20 and Oct. 3, 2007 to collect the information about vegetation coverage, topographic and hydrological conditions across the watershed and particu-

Table 1. Sensitivity Analysis of WATFLOOD Parameters

Adjustment of parameters (one at a time)		↓30%	↓15%	↓5%	↑5%	↑15%	↑30%
Parameter	Explanation	Variation of logarithmic NSE (%)					
AK2	upper zone drainage resistance factor for bare ground	1.4	0.5	-0.1	-0.7	-1.2	-1.9
AK2FS	upper zone drainage resistance factor under snow	0.2	-0.1	-0.3	-0.5	-0.7	-1.1
AK	soil permeability of bare ground (mm/h)	0	0	0	0	0	0
AKFS	soil permeability under snow (mm/h)	0	0	0	0	0	0
Albedo	the all-wave albedo	-0.4	-0.4	-0.4	-0.4	-0.4	-0.4
BASE	base temp for snowmelt (^{o}C)	-37.1	-18.5	-8.8	1.9	7.5	16.7
MF	melt factor (mm/^{o}C/h)	6	5.9	1.7	-2.7	-7.5	-14.9
NMF	negative melt factor (mm/^{o}C/day)	-2.2	-1	-0.7	0	0.5	1.3
R3	overland flow roughness for bare pervious area	-0.4	-0.4	-0.4	-0.4	-0.4	-0.4
R3FS	overland flow roughness for snow covered pervious area	-0.4	-0.4	-0.4	-0.4	-0.4	-0.4
REC	interflow depletion coefficient	0.1	0	0	0	-0.1	-0.3
RETN	upper zone specific retention (mm)	-0.4	-0.3	-0.3	-0.3	-0.3	-0.3
A5	API hourly reduction value	0	0	0	0	0	0
lzf	lower zone drainage function parameter	3.1	1.3	0.4	-0.4	-1.2	-2.2
pwr	lower zone drainage function exponent	35.8	15.3	4.4	-3.9	-11.5	-20.9
R2n	river channel Manning's roughness coefficient	-11.4	-5.8	-1.3	1.2	4.4	7.4
ds	depression storage for bare ground (mm)	-0.4	-0.4	-0.4	-0.4	-0.4	-0.4
dsfs	depression storage for snow covered ground (mm)	-0.4	-0.4	-0.4	-0.4	-0.4	-0.4
flapse	lapse rate in ^{o}C per 100 m (^{o}C)	-0.4	-0.4	-0.4	-0.4	-0.4	-0.4
fpet	increase in interception evaporation for tall vegetation	-0.6	-0.5	-0.4	-0.3	-0.3	-0.2
ftal	reduction in soil evaporation due to tall vegetation	-33.5	-13.3	-4.5	3.3	8.1	12.2
kcond	conductivity of the wetland (mm/h)	0	0	-0.1	-0.7	-1.6	-3.5
mndr	meandering factor	-	-	-	-0.4	-0.4	-0.4
R1n	flood plain Manning's roughness coefficient	-0.4	-0.4	-0.4	-0.4	-0.4	-0.4
sublim	crude snow sublimation factor (mm/h)	-29.7	-11.9	-3.2	-0.1	0.1	0.5
theta	porosity of the wetland or channel	-35.8	-15.5	-4.8	3.4	10.8	19.3

larly in the upper reach of the river. To further validate the findings from field investigation, the WATFLOOD hydrological model was applied to mimic the hydrological processes in the study area.

3.2. The WATFLOOD Hydrological Model

The hydrology of the Deer River watershed was simulated using the conceptual, distributed hydrological model WATFLOOD that has been developed at the University of Waterloo (Tao and Kouwen, 1989; Kouwen et al., 1993). WATFLOOD has been designed and widely used for long term river flow simulation or flood forecasting. The concept of grouped response units (GRU) allows its application to large watersheds where similar vegetated areas within each grid segment are grouped as one land cover type for water balance calculation (Figure 2). The model assumes that similar land covers exist in regions of homogenous soil characteristics and topographic conditions. WATFLOOD simulates both vertical and horizontal water balance in the natural environment, including interception, infiltration, evapotranspiration, snow accumulation and ablation, interflow, recharge, baseflow, and overland and channel routing (Kouwen et al., 1993). Snowmelt is simulated by an index method which allows refreezing (Anderson, 1973). Infiltration process is governed by the Philip formula (Philip, 1954) and Darcy's law. The Priestley-Taylor (Priestley and Taylor, 1972) and Hargreaves' equation (Hargraeves and Samani, 1982) are employed to estimate the potential evapotranspiration based on the data availability. The actual evapotranspiration is either presumed as the potential rate when soil moisture is at a level of saturation or reduced to a fraction of the potential rate if the soil moisture is below the saturation point. Channel flow is routed based on continuity and Manning's formula, whereas base flow is calculated by a non-linear storage-discharge function.

3.3. Modelling Inputs

Spatial data used in this study included a digital elevation model (DEM), land cover, meteorological and streamflow records. A 3-arc-second (90 m resolution) DEM was downloaded from the Seamless Data Distribution System, National Center for Earth Resources Observation and Science (EROS), U.S. Geological Survey (USGS, 2008) and processed with TOpographic PArameteriZation (TOPAZ). Land cover datasets (1 km resolution) 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. Daily streamflow data (1978 to 1997) at the D. River N. Belcher station (ID: 06FD002, Figure 1) were provided by Water Survey Canada.

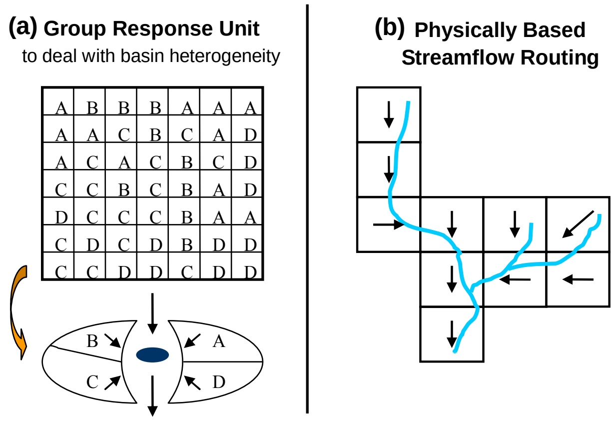

Figure 2. Group Response Unit and runoff routing concept (Donald, 1992).

Hourly meteorological data (1978 to 1997), including air temperature and precipitation were provided by Environment Canada at the nearest station, Churchill-A Climate station (ID 5060600). The degree-day index method was selected for snowmelt calculation when ambient air temperature exceeds the rainfall/snow division temperature. Potential evapotranspiration was estimated using the Hargreaves' equation due to the lack of hourly net radiation data.

3.4. Sensitivity Analysis and Modelling Validation

To quantitatively assess the variation of model outputs to different sources of variation in the input parameters, sensitivity analysis was performed using the one-factor-at-a-time (OFAT) technique. Each parameter was manually adjusted (±5%, ±15%, and ±30% from the initial values) at a time while keeping others fixed. The initial values were determined based on the filed investigation, the WATFLOOD manual and other researcher's work close to the study area (Bellisario et al., 1999; Metcalfe and Buttle, 2001; Shaw et al., 2005). Variation of model outputs was evaluated by the fluctuation of the logarithmic Nash and Sutcliffe efficiency (NSE_{ln}) for a 10-year period (1978 to 1987) as follows:

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

where Q_0, Q_m and $Q_{average}$ are the daily observed flow (m^3/s), the daily simulated flow (m^3/s) and the 10-year mean observed flow (m^3/s) at the D. River N. Belcher station, respectively. Hourly simulated flow was averaged to daily scale in order to compare with the observed records. NSE_{ln} value is less than or equal to 1. The closer it is to 1, the better performance of the WATFLOOD model. The results indicated that lower zone drainage function exponent, base temperature for snowmelt, porosity of the wetland or channel bank, reduction in soil evaporation due to tall vegetation, crude snow sublimation factor, and snowmelt factor are the most influential parameters that should be optimized during modeling calibration (Table 1).

The most influential parameters were calibrated using the first 10-year (1978 to 1987) data to achieve the best model performance. Model performance was measured by NSE_{ln} and the deviation of runoff volumes (D_v) at the D. River N. Belcher station to determine whether the calibration is completed. As also known as the percentage bias, D_v is calculated by:

$$D_v(\%) = \frac{\sum(S_i - O_i)}{\sum O_i} \cdot 100 \quad (2)$$

where S_i and O_i are the daily simulated and observed flows (m^3/s) at the D. River N. Belcher station, respectively. Values of other parameters were fixed as their initial values in reference to the WATFLOOD manual and other researcher's work close to the study site. Modelling verification was performed for the subsequent 10-year period (1988 to 1997).

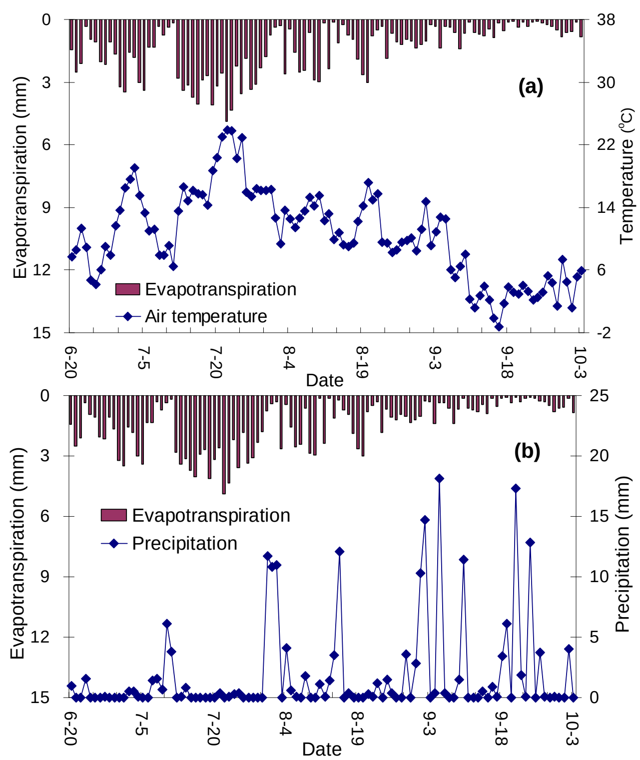

Figure 3. Variation of daily evapotranspiration and (a) air temperature, and (b) precipitation at Rail Spur in 2007.

4. Results and Discussions

4.1. Hydrological Processes in Summer

The 3-year summer observations at Rail Spur showed a significantly proportional relationship between daily air temperature and evapotranspiration in summer; meanwhile, precipitation played an important role in elevating the evapotranspiration with an average lag of one day. For example, daily evapotranspiration reached its local bottom (0.2 mm/day) on July 11, 2007 along with the local minimum daily temperature

(6.5 °C) (Figure 3a). Heavy rainfalls occurred on July 10 and 11 (6.1 and 3.8 mm/day, respectively) along with the gradually increasing temperature; consequently, the daily evapotranspiration rose up to 4.4 mm/day one day after its local lowest point (0.7 mm/day) on July 11 (Figure 3b). This demonstrated that air temperature is one of the dominant factors of summertime evapotranspiration in subarctic wetlands. Meanwhile, precipitation also influenced evapotranspiration process because it increased water availability and air humidity. For example, daily evapotranspiration reached its local bottom (0.2 mm/day) on July 11, 2007. With the gradually increasing temperature, heavy rainfalls occurred on July 10 and 11, 2007 (6.1 and 3.8 mm/day, respectively); consequently, the daily evapotranspiration increased up to 2.8 mm on July 12, 2007 (Figure 3b).

Soil layers became more saturated as getting closer to the stream channels. The average soil moisture contents at the left bank of Station 5 were observed as 38.2%, 24.5%, 21.8% and 17.9% with the locations varying from 2 to 8 m (2 m interval) away from the bank in 2006 (Figure 4a). On account of the presence of extraordinarily high hydraulic conductivity (Reeve et al., 2000), near-stream locations received more infiltrated water from streams and the vicinity because the extent of this infiltration process was inversely proportional to the distance from the bank. Some discrete points that disobeyed this trend could be attributed to the presence of permafrost table, soil texture and land slope. Permafrost table descends as approaching to the stream, which represents that active organic layer become deeper and infiltration occurs more easily with less resistance because the storage capacity increases. Therefore, some distant transects, if permafrost table is shallow enough, are possibly to be saturated near the ground surface, leading to higher soil moisture contents than those close to the stream. Measurements from the automated weather station at Rail Spur also disclosed the temporal and vertical distribution of soil moisture and soil temperature (Figure 4b). Following the major recharge period during the snowmelt, soil moisture contents kept declining throughout the summer, mainly due to the intensive evapotranspiration. In addition, the descending permafrost table enlarged the active organic layer and dragged the water table downwards, which reduced the water supplement. Soil temperature of the shallow layers (at 0, 5 and 10 cm depth) varied continuously and to some extent in accordance with air temperature (Figure 4c), whereas soil temperature of the deep layers (50 and 75 cm) kept stable (around 0 °C) which could be attributed to the fact that permafrost lied around.

A reciprocal relationship between permafrost table and its distance to the stream channels was observed (Figure 5a). There are a number of possible factors which may contribute to this finding, such as the effects of water contents from stream flow and subsurface flow as well as low albedo vegetations. Soil moisture tended to be higher as getting closer to the stream where porous organic layer became saturated from stream penetration and the thaw of ice contents, which in turn accelerated the depression of the permafrost table. This melting process could be intensified by subsurface flow that moved towards the stream. Coniferous forests locating along the streams have much lower albedo than the lichens/moss covered hollows which

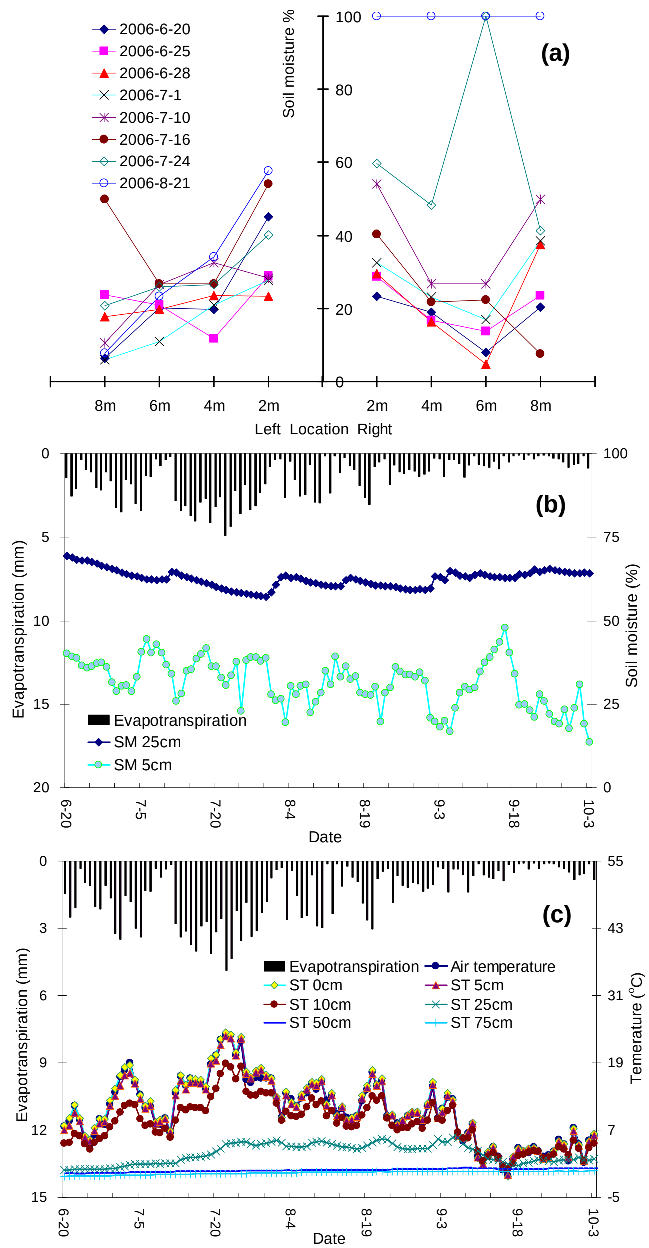

Figure 4. (a) Soil moisture at the banks of station 5 in 2006 summer, (b) variation of daily soil moisture and evapotranspiration in 2007 summer at Rail Spur (SM: Soil Moisture) and (c) multiple soil layers temperature, air temperature and evapotranspiration in 2007 summer at Rail Spur (ST: Soil Temperature).

are far away from the streams. This difference may result in the absorption of more radiation energy and subsequently the aceleration of the ice thaw at the near-stream locations. Similar evidences have also been reported by the previous studies (Woo and Marsh, 2005; Woo and Young, 2006). Air temperature acted as the dominant factor of the fluctuation of permafrost table as shown in Figure 5b. For instance, permafrost table continuously descended throughout the summer of 2007 when air temperature retained at 11 to 23 °C. Meanwhile, the

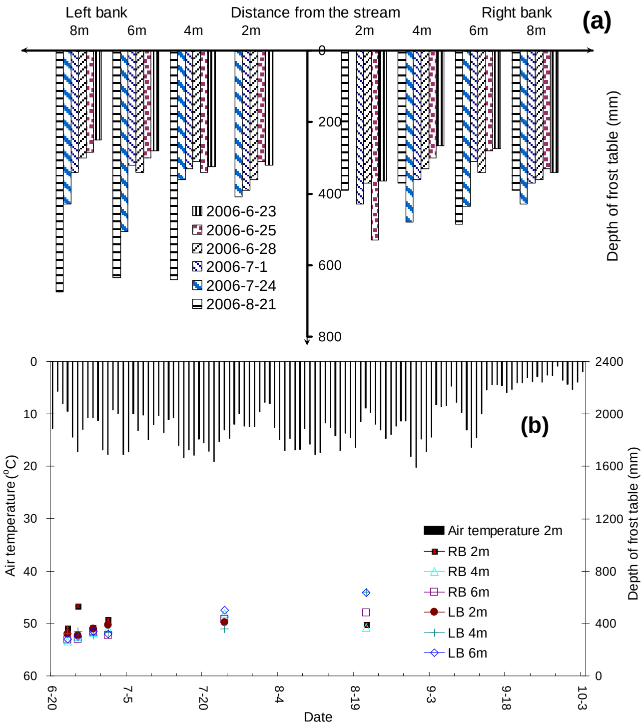

Figure 5. (a) Permafrost table and (b) permafrost table v.s. air temperature in 2006 at station 6 (RB: Right Bank; LB: Left Bank).

influence of precipitation on permafrost table was not as significant as that of air temperature. Permafrost table kept descending regardless rainfall events, which mainly contributed to the fluctuation of groundwater table.

Streamflows had a descending trend before September 2007 and an ascending one thereafter (Figure 6a). Most of the small or moderate rainfall events during the summer barely generated noticeable runoff. This phenomenon was due to the descending permafrost table, high soil porosity and intensive evapotranspi- ration. Rainfall events occurred in the fall often resulted in relatively high volume of runoff because the evapotranspiration was alleviated by low temperature and net radiation. A lag of 1 ~ 2 days between the peaks of streamflows and rainfall events was found at most of the monitoring stations due to the effect of runoff concentration (Figure 6b). These findings were in line with some observations previously reported by other researchers (Quinton and Roulet, 1998; Carey and Woo, 1999; Carey and Woo, 2001; Woo and Marsh, 2005; Woo and Young, 2006).

4.2. Hydrological Modelling

Table 2 and Figure 7 summarize the modelling outputs at the D. River N. Belcher station for both calibration and verification periods. It is indicated that snowmelt in the spring season (April to June) was the major source of water supplement. Usually it produced approximately half of the annual water recharge in subarctic wetlands as shown in Figure 8. Peak of the simulated spring runoff was lower than the observed peak which could be attributed to a number of possible reasons.

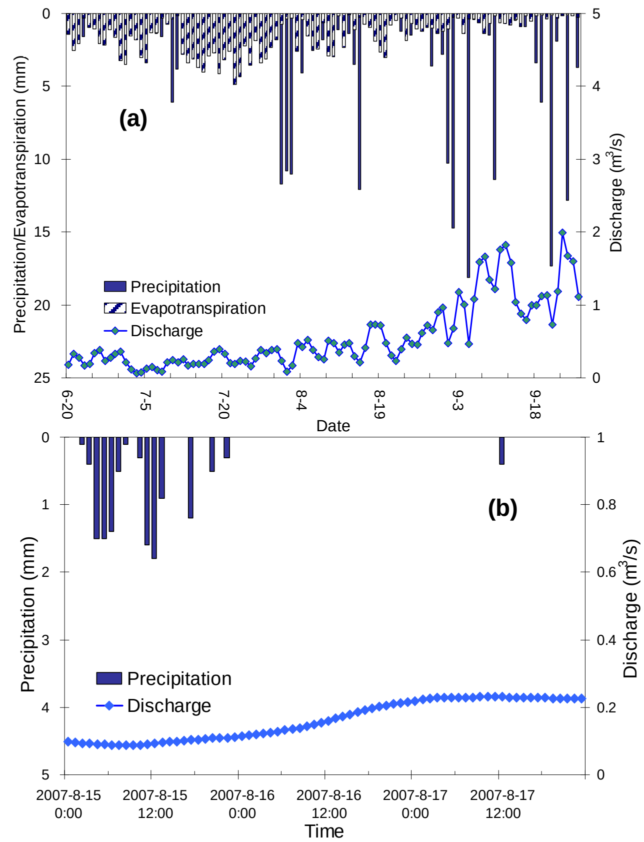

Figure 6. (a) Plot of evapotranspiration, precipitation and water discharge in 2007 summer at station 5 and (b) response of hourly water discharge to precipitation at station 10 (August 15 to 17, 2007).

Permafrost represents an over-winter surface storage of groundwater. When temperature increased and snow started to melt, permapermafrost table began to decline due to the thaw of ice contents within the soil layers. Stream runoff was amplified because the descending permafrost layer could impede the percolation of water. Another explanation to this peak difference is the existence of a large number of seasonal ponds and lakes. These ponds were able to store and release large amount of water and therefore replenished spring runoff when snow and ice started to melt. The modelling results also stated that most of the simulated snowmelt events were 5 ~ 10 days later than the actual melts. This delay can be explained by the fact that permafrost restrains the water infiltration and accelerates the runoff concentration.

Most light and moderate rainfall events in summer (July to September) generated relatively small amounts of runoff which can be related to canopy interception, depression storage, porous soil layers, impermeable permafrost and intensive evapotranspiration. As observed during field investigation, many lichens, mosses, small shrubs and conifers flourish in the summer. A relatively great amount of precipitation during light rainfall events could be intercepted by their canopy storage. This amount of water could be evaporated back to the atmosphere afterwards. Depression storage is mainly referred to the numerous seasonal ponds that stretched over the study area. A

Table 2. Summary of Annual Modelling Outputs at the D. River N. Belcher Station for the Deer River Watershed (1978-1997)*

Year	Cumulative P (mm)	Cumulative simulated ET (mm)	Mean T (^{o}C)	Mean observed streamflow (m^3/s)	Mean simulated streamflow (m^3/s)	NSEln (%)	DV (%)
1978	532.8	403.9	-7.76	12.7	24.5	8	93
1979	341.9	204.4	-7.96	13.9	17	41	23
1980	484	332.2	-6.93	18.4	13.7	66	-31
1981	395.4	325.2	-4.89	14.4	22.5	38	56
1982	605.8	527.3	-8.43	15.8	35	20	93
1983	621	610.2	-7.06	27.9	49.6	36	78
1984	413.9	235.7	-6.5	14.3	26.5	20	86
1985	448.5	325.4	-7.11	11.1	15.4	28	39
1986	500.3	431.4	-7.33	22.8	30.8	49	35
1987	432.7	299.6	-5.42	11.6	22.5	25	81
1988	441.3	207.9	-7.38	9.07	13.3	48	46
1989	358.5	232.9	-8.02	17.6	16.6	58	-6
1990	485.3	320.4	-7.41	12.1	15.6	41	29
1991	524.9	419.9	-7.09	23.8	24.5	55	3
1992	402	401.8	-8.21	12.8	26.4	13	106
1993	291.5	135.1	-7.11	9	9.9	25	10
1994	345.5	186.3	-6.62	7.2	6	52	-16
1995	416.4	397.3	-6.9	19.4	22.4	52	15
1996	424.1	233.1	-7.46	7	16.9	7	142
1997	509.7	462.5	-6.55	27.2	27.8	68	2
Average	448.8	334.6	-7.11	15.4	21.8	38	44

* P is precipitation; ET is evapotranspiration; and T is temperature.

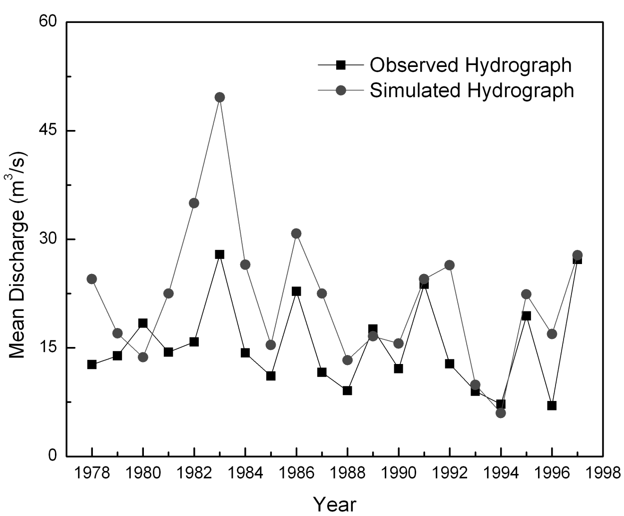

Figure 7. Temporal trends in simulated and observed annual hydrograph at the D. River N. Belcher station.

great number of seasonal ponds were observed in the study area while they can store a great amount of precipitation and prolong the runoff concentration. The hummocky terrain prevails in the study area and consists of porous peat overlying a thick layer of mineral substrate. This particular structure of the soil layers along with the descending permafrost allows a great amount of water to be temporarily stored. Precipitation is retained in soil layers and ponds for a longer period and more likely to be released as evapotranspiration or groundwater flow rather than surface runoff. As depicted by the analysis of meteorological records, there is usually a pronounced peak of evapotranspiration in summer due to higher air temperature and longer daylight period. Increased evapotranspiration brought water back to the atmosphere and therefore reduced 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 fall generated much more runoff due to relatively lower temperature and less net radiation.

A lag of 2 to 8 days between the peaks of rainfall and stream runoff was observed in both summer and fall. As shown in Figure 9, a short-duration (38 hours) and high-intensity (65 mm in total) rainfall event occurred on September 2 and 3, 1983. The simulated and observed runoff peaks were delayed until September 5 which indicated that the runoff concentration was prolonged to around 2 days. For heavy rainfall events, after the surface soil layer getting saturated, excess water generated flashy runoff and resulted in steeper runoff response. Contrastingly, concentrated but moderate rainfall events tended to have longer runoff concentration time because of the large buffering capacity of wetland water storage, which also validated the conclusions from the investigation. Water contents of the wetland system, including groundwater storage and surface runoff drastically decreased due to the intensive evapotranspiration in summer and fall. For example, during the continuous rainfall event from September 18 to 21, infiltration dominated water distribution and brought most of the precipitation into soil layers. Eventually the peaks of simulated and observed runoff appeared 5 days later with a prolonged and gentler runoff response.

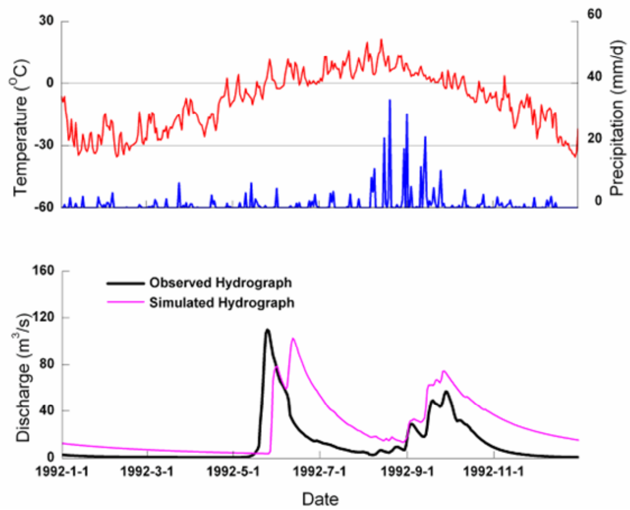

Figure 8. Simulated and observed daily hydrographs for the Deer River Watershed in 1992.

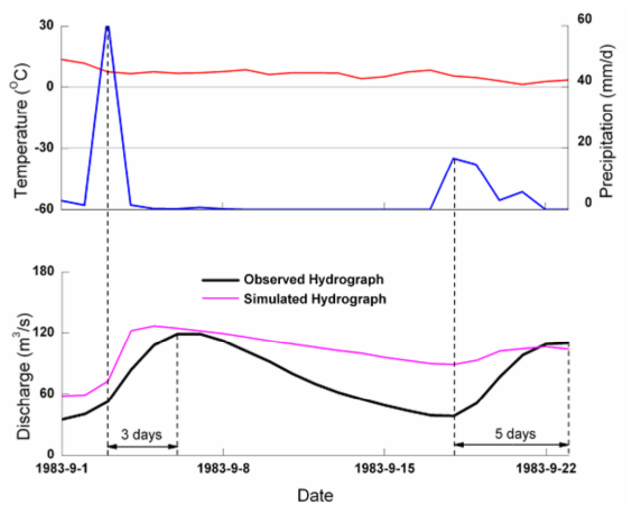

Figure 9. Response of daily discharge to precipitation by the WATFLOOD model in 1983.

Although the modelling results reasonably matched the observed data, some limitations should be noticed. Firstly, the WATFLOOD model was targeted and tested during the period between 1978 and 1997 due to limited data availability when the modelling work was conducted. It is likely that regional and global conditions would have drastically changed such that the representativeness of the modelling outputs may be impaired. Nonetheless, outputs from this study are still valuable in examining the feasibility of the WATFLOOD model in simulating subarctic watersheds and demonstrating the unique hydrologic features of subarctic wetlands in northern Manitoba. Secondly, 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. Thirdly, 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). Lastly, WATFLOOD has distinguished sub-routines that take account of snow sublimation and wetland water distribution. However, parameters of these sub-routines were not studied during the investigation and their values were optimized during the model calibration.

5. Conclusions

This research presents an integrated study of the hydrology of subarctic wetlands through field investigation and hydrological modelling. An extensive field investigation of the Deer River watershed near Churchill, Manitoba was conducted in the summertime from 2006 to 2008. A monitoring network was established to collect hydrological and meteorological data for the in-depth understanding of the sub-arctic wetland attributes. Air temperature appeared to be the primary driving force for wetland evapotranspiration while precipitation had limited influences. Surface soil moisture became more saturated as getting closer to the stream which could be attributed to the extraordinarily high hydraulic conductivity and the descending permafrost. Permafrost table kept descending in the summer and released extra frozen soil layers due to the increasing temperature. In addition, low albedo vegetation and subsurface flow determined a reciprocal relationship between permafrost table and its distance to the streams. These findings could explain the fact that summer rainfall events were not able to generate noticeable runoff. To quantitatively confirm these findings, a conceptual distributed hydrological model (WATFLOOD) was employed to simulate the hydrological features of the Deer River watershed. Sensitivity analysis (one-factor-at-a-time) was conducted to determine the target parameters for model calibration. Model veryfycation was performed based on the values optimized by model calibration. The results indicated that snow-melt usually produced the highest peak flow and the majority of annual runoff. The existence of permafrost reduced the concentration time of spring runoff due to its impermeable attributes. Summer rainfall-runoff response was also greatly influenced by the declining permafrost which activated frozen soil and prolonged the runoff concentration. Regardless of the uncertainties caused by low resolution input and unmeasured model parameters, the application of the WATFLOOD model demonstrated a number of particular features of subarctic wetlands.

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