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Hydrological Response of a Patchy High Arctic Wetland

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High Arctic patchy wetlands are ecological oases in a polar desert environment and are vulnerable to climatic warming. At present, understanding of their responses to external factors (climate and terrain) is limited. This study examines a wetland located in a topographic depression maintained by seasonal snowmelt, ground ice melt and lateral inflows. The wetland is located on Cornwallis Island, Nunavut, Canada. Hydrological, climatological and soil observations were made over several summers with different weather conditions. The summers of 1996 and 1997 were cool and wet but the summer of 1998 was warm and dry. The melt in 1996 was rapid due to rain on snow events and only lasted six days. Deeper snow in 1997 prolonged the melt season to 18 days. A shallow snow-cover in 1998 and early melt depleted the snow by early June. Surface, ground-water and storage fluctuations in the wetland were dictated by snowmelt, rainfall, evaporation loss from the wetland and lateral inputs which in turn were controlled by the melting of the late-lying snow storage in the catchment. Soil factors influence the spatial variations in ground thaw which affects the surface and subsurface flow. Streamflow response of the wetland reflects a nival regime and augmentation of streamflow thoughout the summer season in all three years is supported by multiple water sources: ground ice melt and suprapermafrost water from a large late-lying snowpack. Overall, this study suggests that the survival of some patchy wetlands depends on their interaction with the surrounding basin, with a dependency probably being more important during warm and dry seasons.

Introduction

Wetlands comprise a large area of northern Canada and are signficant ecological environments. Clearer understanding of their present day hydrological function is required before we can anticipate their role in uptaking or releasing greenhouse gases, and their function in maintaining and sustaining plant biodiversity. In the Canadian High Arctic, large extensive wetlands do exist but usually in "polar oasis" type environments which enjoy warmer temperatures and greater radiation than the majority of the High Arctic which can be considered to have a "polar desert" type environment (cool temperatures and low radiation levels) (Woo and Young 1997). Aside from these lush vegetated areas which have been examined to some extent (*i.e.* Rovansek *et al.* 1996; Ryden 1977), less work has focussed on small patchy wetlands which may exist in breaks of slopes, along riverine shores, ponds and in small topographic depressions where water seepage may occur. Glenn and Woo (1997) examine a small valley-bottom wetland on the Fosheim peninsula (polar-oasis type environment) and Woo and Young (1998) describe the water table regime of several patchy wetland types in a polar desert environment. In the Low Arctic, Roulet and Woo (1986,1988) describe the hydrology of a small wetland modified by a lake. Several of these studies (*i.e.* Glenn and Woo 1997; Roulet and Woo 1986, 1988) suggest that streamflow response from these wetlands is limited to the snowmelt period and the surrounding landscape does not play an important role in water flow through the wetland for a prolonged period. In some cases, observations have been made only for a single year (*i.e.* Glenn and Woo 1997; Woo and Young 1998) and therefore it is difficult to assess how these patchy wetlands will respond under variable climate conditions. This study was undertaken to examine the hydrological response of a patchy wetland in the polar desert over several summer seasons which varied in terms of temperature and rainfall patterns. This study will clarify how high arctic patchy wetlands interact with the surrounding basin and how they respond to changing climatic conditions.

Study Area

The field study was conducted near Resolute Bay, on the S.W. coast of Cornwallis Island (74°15'N, 76°W) from mid-June to Aug. 1996 (Fig. 1). The area is underlain by continuous permafrost and active layer depths range from 0.3-1.0 m depending on vegetation cover. The area has rolling topography with elevations between sea level and 200 m (Woo and Steer 1983). Most of the ground surface is barren with vegetation growing at wet sites. Less than 1% of the island is occupied by wetlands. The regional climate is characterized by low air temperatures, short cool summers and long cold winters. Annual precipitation, when corrected for snow gauge undercatch, is about 200 mm about 80 per cent of which comes as snowfall (Woo *et al.* 1983). The snow cover at the end of winter is uneven due to the treeless terrain and

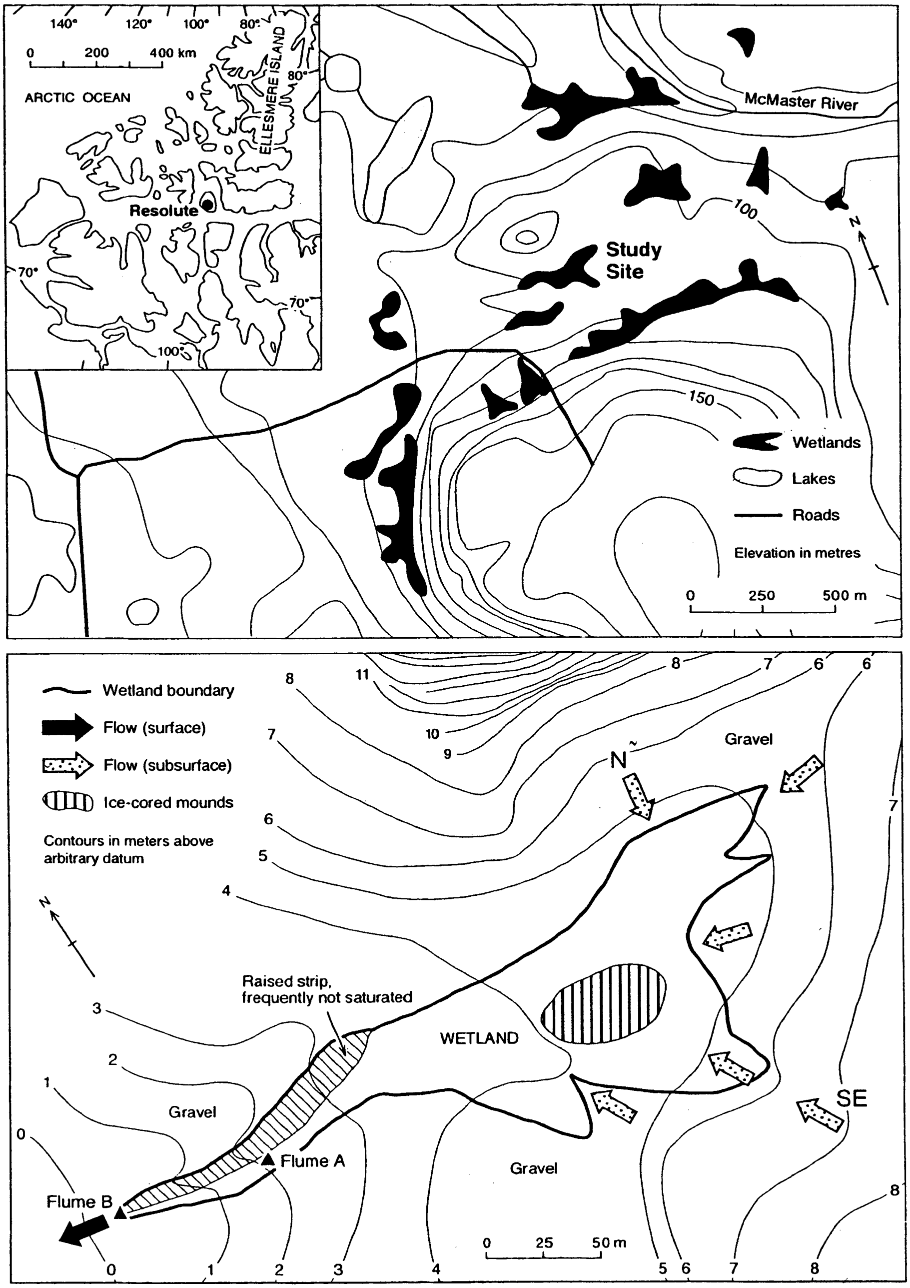

Fig.1. Study area location in the Canadian High Arctic (a), and its relationship to other patchy wetlands in the McMaster River Basin (b). The wetland boundary, topography and flow configuration are found in part (c).

Table 1 – Percentage cover characteristics of multi-source wetland based on quadrat surveys

	Wetland Area (%)	Non-wetland Area (%)
Gravels	2.9	40.4
Mud and Litter	7.8	11.6
Water	7.8	0.0
Algae	9.2	20.8
Moss	45.8	16.1
Lichen	0.8	2.2
Vascular Plants	25.7	8.9

Note: Vegetation was identified along 3-4 transects across the wetland and non-wetland basin and the plant coverage was estimated at 2.5 to 5 m intervals from quadrats measuring 0.2 mx0.5 m or 0.5 mx0.5 m depending on plant cover (Batten and Svoboda 1994).

Table 2 – Vascular plant species found in multi-source wetland and ranked importance values ( ).

Species	
Alopecurus alpinus (1)	*Eriophorum scheuchzeri*
Agropyron latiglume var. pilosiglume (4)	*Eriophorum triste*
Agropyron violacium var. hyperarcticum	*Juncus biglumis (3)*
Carex membranacea	*Gramineae sp.*
Carex misandra	*Papaver radicatum*
Cerasitum arcticum	*Pleuropogon sabinei*
Cerasitum alpinum (2)	*Ranunculus hyperboreus*
Cerastium regelii	*Saxifraga cernua*
Draba alpina	*Saxifraga oppositifolia*
Draba lactea	*Trisetum triflora sp. inoble*
Draba sp.	

Note: Vegetation survey data also allowed importance values of vascular plants to be determined. It is defined as the product of the % of plots in which each species occurred and the average cover value of each species in those plots in which it occurred (Gorham *et al.* 1998).

Table 3 – Ground ice content in the active layer with depth and near surface bulk density, porosity and organic content after Woo and Young (1998).

Depth (m)	Ground ice Content (%)	Avg. Bulk Density (kg/m3) 0-0.15 m	Avg. Bulk Density (kg/m3) 0-0.15 m	Avg. Porosity (%) 0-0.15 m	Avg. Porosity (%) 0-0.15 m	Organic Content (%) 0-0.15 m	Organic Content (%) 0-0.15 m
0.03	83	Wetland	Non-wetland	Wetland	Non-wetland	Wetland	Non-wetland
0.10	33	165 n=11	250 n=19	37.6	12.2	30.4 n=16	3.9 n=10
0.17	42						
0.27	39						

strong northwesterly winds which are effective in redistributing snow. Snowmelt usually begins in June and ground thaw continues until late August. The mean annual temperature is -16.6°C, and only the months of July and August have mean temperatures above freezing (Washburn 1997). The surface materials of Cornwallis Island are largely derived from local bedrocks which are frost shattered due to the harsh northern climate. These materials can influence the infiltration of meltwater into the soil. Beneath the surface, Paleozoic limestone and dolomite dominate the Island; however, calcareous shale, siltstone, gypsum, anhydrite and sandstone are also found (Thorsteinsson 1958; Thorsteinsson and Kerr 1968). More specifically, Resolute is underlain by Silurian limestone of the Read Bay Formation (Thorsteinsson 1958).

The multi-source wetland (see Fig. 1b,c) receives water from a distant late-lying snowbank located on a concave slope and from groundwater seepage. The wetland area is approximately 28,800 m^2 and has a six- metre elevation range, giving it an average slope of 1.7° (Stoltz 1998). This site is undulating and the topography forms a depression in the centre where organic matter accumulate. It is also characterized by ice cored peaty mounds and raised peaty zones which remain dry. Tables 1 and 2 indicate the percentage cover characteristics for the wetland and its biodiversity. Soil properties of near-surface soils and ground ice content with depth are provided in Table 3. Within the wetland, 84% of the near-surface materials are less than 180 mm (n=13) in size, while in the non-wetland gravel zone, this falls to 36% (n=10). The type of cover, soil properties and ice content can influence subsurface water flow and the thaw rate of the active layer.

Methods

Interactions among the atmosphere, the soils and the frozen substrate and lateral water movement govern the hydrology of northern wetlands. For a wetland to be sustained, a positive water storage must be maintained frequently and hence, water inputs into the system must exceed outputs (*i.e.* precipitation, snowmelt and lateral flow must be larger than evaporation and flow leaving the wetland)

$$\Delta s = M + P + Q_g + Q - E \tag{1}$$

here Δs is the change in storage in the saturated and unsaturated zones (m/d); M is snow or ice melt (m/d); P is precipitation (m/d); Q_g is net groundwater flow, considered to be the difference between the input and output; Q is net surface runoff (m/d); and E is evaporation (m/d). The change in storage of a wetland on a daily or yearly basis increases or decreases according to the amount of inputs and outputs. The water table being the upper limit of the saturated zone, dictates whether or not plant roots have easy access to water and nutrients, and therefore influences the evaporation pattern. It also defines the zones in the wetland where decomposition or accumulation will occur, thus whether a wetland will be a source or sink for CO_2 and other greenhouse gases (Rouse *et al.* 1997). The water table was monitored manually on a daily basis in six transects of perforated and screened pipes which traversed the wetland and the surrounding non-wetland basin. Frost table was determined by pounding an iron rod into the ground within a one metre circumference of each water pipe location until it reached the frozen ground (Woo and Xia 1995). This procedure was performed bi-weekly in the early snowfree season following the termination of melt and then weekly throughout the summer season.

The unsaturated zone, which is defined by the soil moisture, was determined on a weekly basis in the vicinity of all water pipes, and bi-weekly at the middle transect within the wetland. No soil moisture profiles were completed but instead near surface soil samples (0-15 mm)(rooting zone) were taken, dried and the gravimetric water content (%) was determined. These measurements together with bulk density measurements as described by Hillel (1971), allowed volumetric water contents (%) to be derived. In addition, several moss samples (125 cm^3) were obtained from the wetland and dried to provide reliable bulk density estimates. The organic fraction of soil was determined by loss on ignition as outlined by Atkinson *et al.* (1958). Average porosity ϕ of the wetland soil was calculated by

$$\phi = \frac{1-D_b}{D_p} 100 \qquad (2)$$

where D_b is the average bulk density of the near surface soils (kg/m^3) and D_p is particle density using the value of 2,650 kg/m^3 (Hausenbuiller 1978). The average storage of the wetland was estimated weekly using average frost table depths, water table depths, porosity and soil moisture estimates for the entire wetland. Snow storage was determined at the end of the winter period in 1996 and 1997 but not in 1998 because of an advanced melt season. Snow cover information was obtained at the beginning of the melt season by placing a steel probe through the snow to the frozen ground surface and measuring the depth. The measuring interval along each transect (5 in 1996; and 6 in 1997) which dissected the wetland and the surrounding basin was two metres and snow density was taken every 10 m using a Canadian MSC snow sampler (Woo 1998). Snowmelt was calculated by an energy balance approach (Young *et al.* 1997) using data collected at a meteorogical tower with sensors set at 1 m above the snow surface. Ouput from the sensors was recorded by a Campbell Scientific CR10 datalogger and measurements were made every minute and averaged over the hour. In addition, daily snowmelt was determined by measuring surface snow lowering at 10 points and the surface snow density (5 measurements daily) using the approach outlined by Heron and Woo (1978).

Here M is computed by

$$M = \Delta d\left(\frac{\rho_s}{\rho_w}\right) \qquad (3)$$

where M is daily snowmelt (m/d), Δd is the amount of snow lowering (m); ρ_s is the surface snow density (kg/m^3) and ρ_w is the density of water (1,000 kg/m^3).

Ground ice melt (m/d) is determined by

$$(FT_{t+1} - FT_t)\ \rho_i\ \frac{\phi}{100} \tag{4}$$

where (FT_{t+1}-FT_t) is the frost table difference (m) between successive days (t and t+1), ρ_i is the ice density (900 kg/m^3), and ϕ is soil porosity (see Eq. (2)). Summer precipitation, P, was measured using a manual raingauge and a recording tipping-bucket raingauge (each tip recorded 0.254 mm of precipitataion). The manual raingauge had a radius of 100 mm and was emptied after each rain event. Trace events were given a value of 0.03 mm.

Inputs of groundwater flow, Q_g, was evaluated along the North and Southeast edges of the wetland (see Fig. 1c) using the Darcy equation

$$Q_g = K_h d_s \left(\frac{dh}{dx}\right) \tag{5}$$

where K_h is the hydraulic conductivity (m/d) and was determined using a pumping test described by Luthin (1966). For the North edge, K_h = 0.033 m/d (n=1) was used and for the Southeast an average K_h of 68.5 m/d (n=2). The term dh/dx is the hydraulic gradient estimated by the elevation of the water table in adjacent wells, and d_s is the thickness of the saturated zone, obtained daily as the elevation difference between the frost and the water tables. To obtain daily frost table estimates for the wetland, a line of best fit was drawn through the weekly frost table measurements. A similar approach was used to determine groundwater flow out of the wetland from the gravel zone between Flumes A and B (see Fig. 1c). However, since reliable hydraulic conductivity measurements could not be obtained in the field at this zone because the recharge was too fast for the pumping test, a hydraulic conductivity of 150 m/d for coarse re-packed gravels was chosen (Morris and Johnson 1967 as found in Todd 1980, pg. 71).

Two flumes and water level recorders were placed near the outlet of the wetland (see Fig. 1c) to monitor surface runoff. The basin discharge, Q, was computed from daily measurements at the flumes where stage levels were monitored electronically with a float hooked up to a potentiometer. Stage records were converted to discharge using empirically determined rating curves. Snowmelt runoff directly coming from the wetland was captured in 1997 only; in 1996, the flume was not established until the last few days of the snowmelt period, and in 1998, due to an early spring the snowmelt runoff from the wetland was over prior to our arrival in the field.

Evaporation was calculated hourly using the Priestley-Taylor (Woo and Rowsell 1933; Lafleur and Rouse 1988; Ohmura 1982) approach and summed over daily intervals. It is given as

$$E = \frac{\alpha(Q^* - Q_c)\,\sigma/(\sigma+\gamma)}{\rho_w L_v} \tag{6}$$

where Q^* is net radiation, Q_c is the ground heat flux, σ is the slope of the saturation vapour pressure *versus* temperature curve, γ is the psychrometric constant, and L_v is the latent heat of vapourization. For open water and freely-transpiring vegetation, the average value of α is approximately 1.26, but in this study the Bowen ratio approach (Lafleur and Rouse 1988; Ohmura 1982; Woo and Rowsell 1993) was used to derive α values. A median value of 0.83 was used in 1996 while in 1997, a median value of 0.66 was applied. Price (1991) also found that α can fall below the equilibrium rate in a coastal marine environment. In 1998, lacking two levels of temperature and vapour pressure measurements, the 1996 value of α= 0.83 was applied (similar climatic conditions (Table 4)). Radiation and all temperature measurements were recorded at the meteorological tower and were averaged hourly. Thermal conductivity of the wetland soil used in the determination of Q_c follows from DeVries (1963).

Results

General Summer Conditions

Among the three field seasons, 1997 had the greatest amount of snow, experienced the coolest temperatures and was wetter over the summer than the other two years (Table 4). The warmest and driest summer was in 1998, reaching an average temperature twice as high as that of 1997, but much drier by a factor of four. The 1996 season was comparable to 1998 in terms of precipitation but was cooler. These data indicate the variability in climate that patchy wetlands in high arctic polar desert environments can experience during the growing season.

Water Sources

Snowmelt and Summer Precipitation – Fig. 2 compares the time of peak snowmelt each day with the occurrence of peak discharge from the wetland. As in other studies (Glenn and Woo 1997; Woo 1988) snowmelt is the dominant water source for the wetland and snowmelt proceeds for several days prior to the initiation of wetland outflow. A time lag occurs as the snowpack needs to become isothermal, and snow drifts caused by the uneven snow distribution may delay water transient or re-route it away from the main channel. Overall, peak snowmelt in the wetland generally occurs between noon and the early evening (1900 hrs), following from daily peak melt energy inputs. The response from the wetland to this water input is generally rapid, occurring a few hours later. This suggests that there is little storage capacity occurring in the wetland at this time as the ground surface remains frozen and the lag is generally governed by the time needed to satisfy depression storage and by the tran-

Table 4 - Conditions during the summer field seasons under investigation

Year	Snow Storage-Wetland (mm)	Average Air Temperature (°C)	Total Summer Precipitation (mm)
1996	126	2.54	32
1997	161	1.95	100
1998	N/A	5.5	22

N/A-Not Available

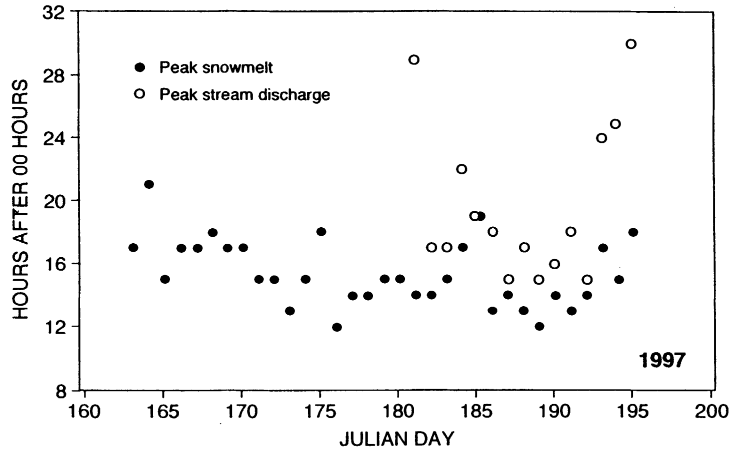

Fig.2. Example of the peak snowmelt and peak stream discharge from the study wetland in 1997.

sient time of the surface flow (Woo 1988). On occasion, peak discharge is delayed until later for two reasons. Early in the season, the melt rate is small (*i.e.* for J.D. 181 M=1.4 mm/h) and travel time is long. Later in the season, despite a high melt rate (*i.e.* for J.D. 193-195, M >5.0 mm/h), the wetland snowpack is disappearing and more meltwater may be seeping into thawed ground and/or travelling from the surrounding basin, hence travel time increases.

In addition to snowmelt inputs, water inputs from summer precipitation can comprise both rain and snow events. Fig.3 provides an example of the the stream response after the precipitation event of July 30-31, 1996 (duration of seven hours) which reached a peak intensity of 2.4 mm/h. The rising limb of the storm hydrograph rose rapidly and the peak occurred approximately seven hours after the centroid of the storm. The lag controlled by depression storage and the transient time of surface flow was only one hour and such a quick response suggests limited storage capacity of the wetland. Following the cessation of the storm, the recession limb falls off sharply. This suggests that even in the summer time when the active layer is

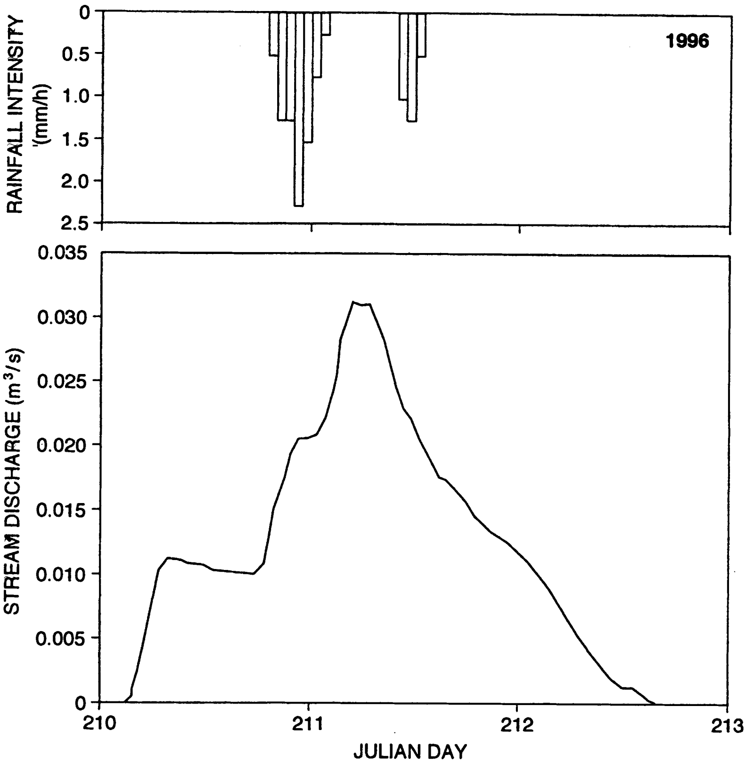

Fig.3. Wetland streamflow response to precipitation inputs, 1996.

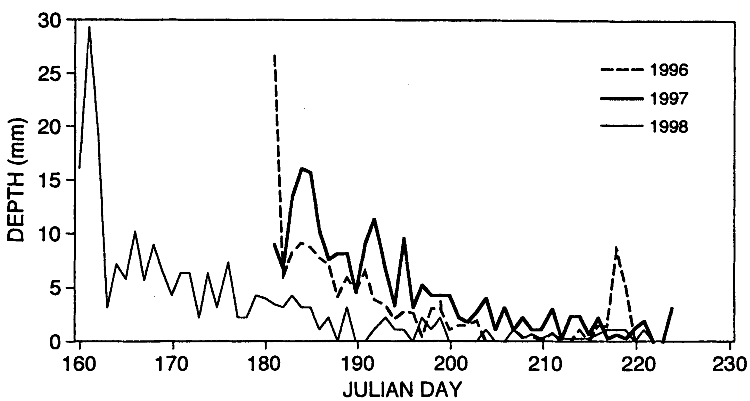

Fig.4. Daily contribution of ground ice melt in the wetland over three summer field seasons (1996-1998).

thawed, the wetland offers poor attenuation of flow. Thus like many arctic wetlands (Woo 1988), the flow does not follow a "muskeg" regime as proposed by Church (1974).

Ground Ice Melt – Fig. 4 indicates the contribution of ground ice melt from the wetland over the three summer seasons. Initially, ground ice melt was large due to rapid thaw, which generally takes place following snowmelt (Young *et al.* 1997). The upper wetland surface also has high ice content (see Table 3) owing to large porosities. Following the release of this water, ground ice melt dropped off and generally yielded less than 5 mm/d of water. This can be attributed to lower soil porosities and ground ice contents in the soil profile (see Table 3) and declining ground thaw rates. This diagram also indicates the variation in the duration and magnitude of ground ice melt which can occur under these three summer field seasons. Due to an early snowmelt and warm summer the wetland in 1998 provided more ground ice melt (23 mm more than 1997; and 80 mm more than 1996). In 1998, the ground ice melt contributions lasted about 20 days longer than the other two previous years. However, in 1997 late season contributions were higher than the other two years.

Water Inflow – Fig. 5 shows the groundwater contribution to the wetland and its surrounding areas. In all years, the contribution from the N section of the wetland (Fig. 1c) is much lower than that from the SE. On the N side, water flow is augmented by lingering snowpatches and ground ice melt, while larger flows on the SE side can be attributed to meltwaters from a large late-lying snowbank (see Fig. 1b and Woo and Young 1998). Meltwater from the snowbank initially flows into a wetland ajacent to it, and a portion of this water seeps into a gravelly zone which separates it from the wetland in this study. Owing to much warmer conditions in 1998, the groundwater flow in this zone of the wetland (SE) began earlier than the other years and rarely fell below 0.05 m^2/d per unit width. On the N side of the wetland, flow was much reduced, and generally constant throughout the summer with slight increases occurring as the frost table deepened. Groundwater flow in this vicinity in 1997 was low (0.001 to about 0.0003 m^2/d per unit width) owing to shallow thaw brought about by cool conditions, while in 1998, flow was higher by about three times.

The receipt of sustained flow to the study wetland from adjacent non-wetland zones can be compared with wetlands elsewhere. Glenn and Woo (1977) found that after the main snowmelt period, the linkages between slopes and a patchy valley fen become detached and the surrounding landscape did not play an integral role in the wetland hydrological processes during the summer period. Bay (1969) also found for a bog watershed in Northern Minnesota, that following the snowmelt period, this wetland type become detached from the surrounding landscape and runoff ceased. Roulet and Woo (1986, 1988) found for a low arctic wetland fed by a lake, after snowmelt, the importance of the wetland in contributing to surface runoff dropped off and water losses from groundwater became more important.

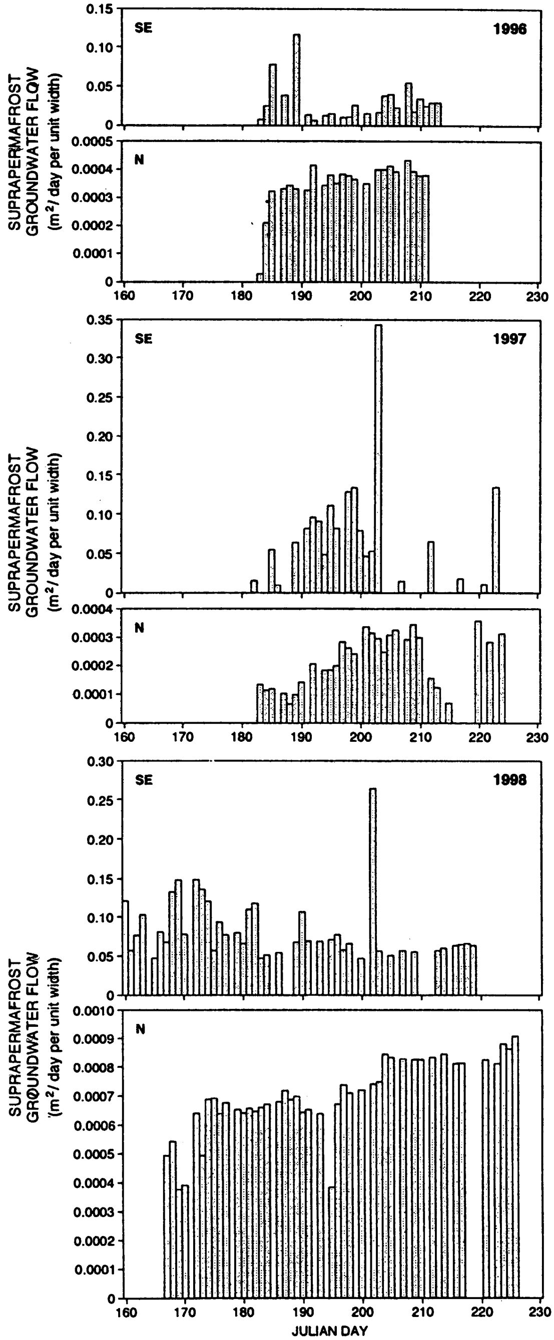

Fig.5. Suprapermafrost groundwater flow into the wetland from different sectors of the basin, 1996-1998.

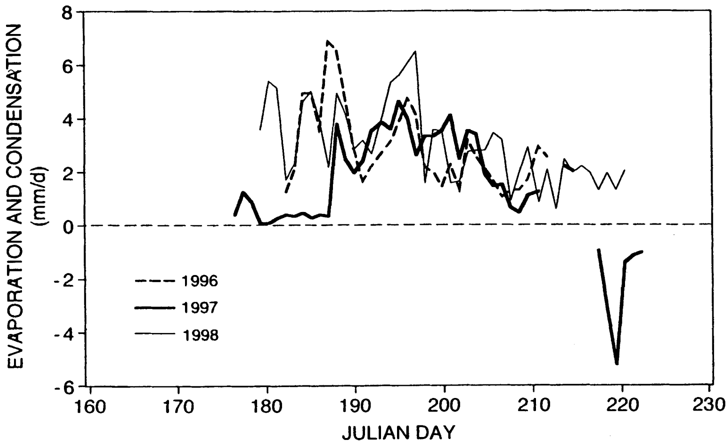

Fig.6. Daily calculated evaporation from the wetland during three summer seasons (1996-1998). Condensation is indicated by negative values.

Water Losses

Evaporation – Evaporation was higher in both 1996 and 1998 owing to more favourable conditions than 1997 (Table 4, Fig. 6). In all years, evaporation was high following the snowmelt period (saturated conditions, high energy levels) and then declined toward the end of the summer (lower energy levels). For example, in 1998 evaporation was about 4.5 mm/d in late June and then dropped to about 2 mm/day by mid-August. Occasionally, condensation becomes important (*i.e.* 1997). The daily evaporation rates were on average 2.9 mm/d in 1996; 1.4 mm in 1997 and 3.0 mm in 1998. This seasonal regime and the evaporation values fall within the range observed by other researchers for both high arctic (Glenn and Woo 1997; Ryden 1977; Woo and Marsh 1990) and low arctic wetlands (Rovansek *et al.* 1996).

Outflow – a) Surface Runoff

Fig. 7 indicates the daily runoff of the wetland from Flume A (see Fig. 1c) which had a more complete discharge record than Flume B over the three year period. Only 1997 adequately demonstrates the wetland's response to inputs of snowmelt from within the wetland itself. Generally, surface flow is delayed beyond snowmelt initiation owing to the time taken to saturate the snowpack, to satisfy detention and depression storage (such as small pools, channels) and the time taken to overcome channel blockages and diversion of water by the uneven snow drifts. Maximum discharge was attained when most of the runoff generating areas became fully integrated. The diurnal rhythm of the hydrographs resulted from inputs of snowmelt from both within (1997) and beyond the wetland boundary from the surrounding basin (1996-1998), including the large late-lying snowbank which fed the study wetland through suprapermafrost subsurface flow in the gravels.

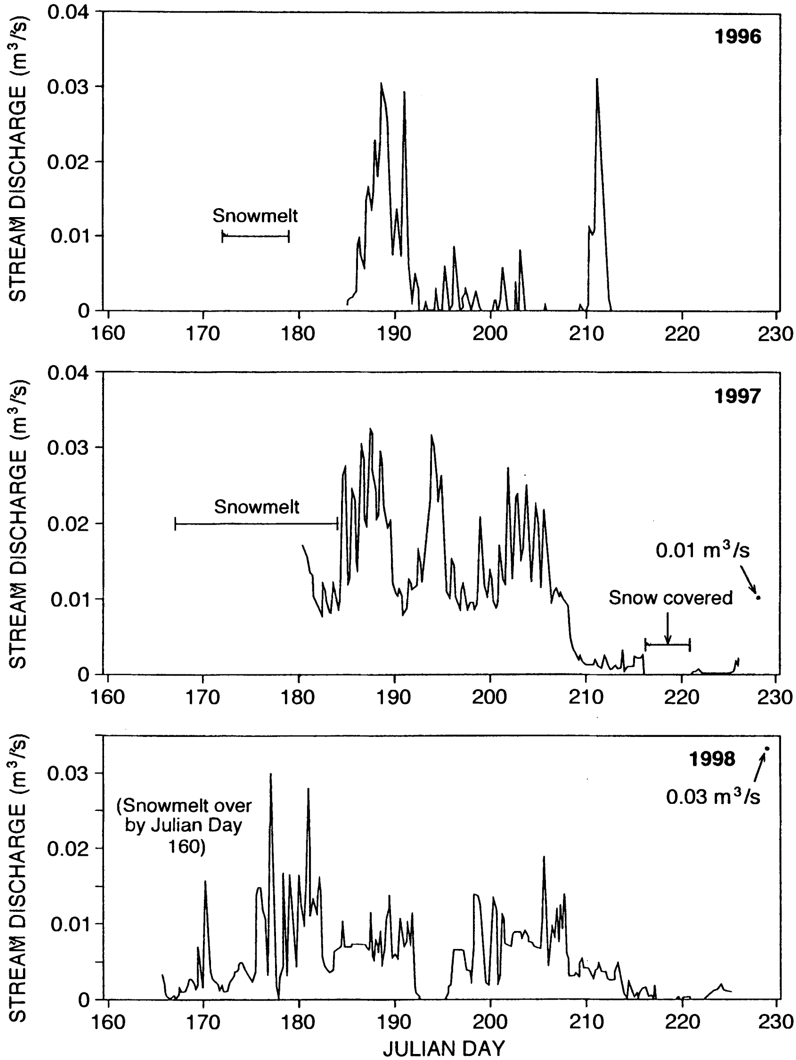

Fig.7. Wetland streamflow response for three summer seasons, 1996-1998. Note that the main snowmelt periods in the wetland are indicated.

While most wetland studies indicate that streamflow ceases following the main snowmelt period and does not rejuvenate (Bay 1969; Glenn and Woo 1997; Rovansek *et al.* 1996; Woo 1988), this study indicates that inputs of water from the areas surrounding the wetland can effectively prolong the streamflow for most of the summer season. Short intervals of low flow to no flow do occur, when the storage capacity increases as evaporation lowers the water levels and the wetland zones be-

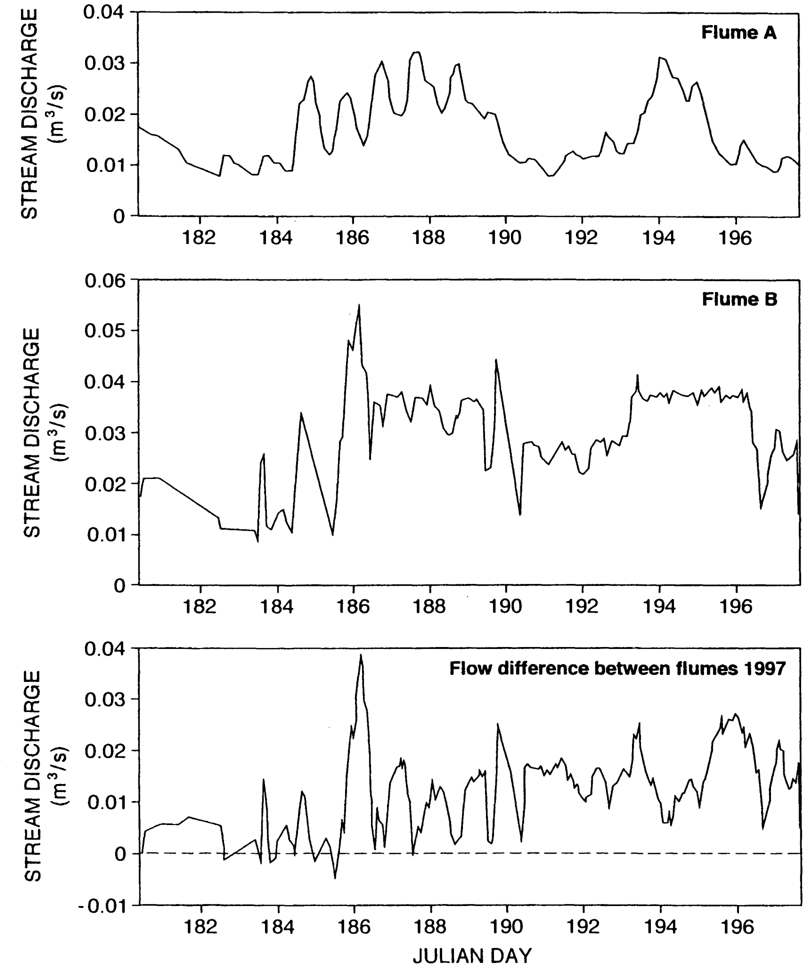

Fig.8. Pattern of surface flow from Flumes A and B in 1997. The difference in flow between the two flumes indicates the water contribution from the northern side of the wetland in comparison to the water supply from the southern side.

come disconnected. Periods of significant snowfall can also effectively shut down the wetland system (see 1997 hydrograph).

In all years, the hydrographs for 1996 (J.D. 210) (also Fig. 3), 1997 (J.D. 229) and 1998 (J.D. 229) show that large summer precipitation events, both rain and summer snow, can generate runoff rates similar to the early snowmelt period. Woo (1998) remarked that small rainfall events after a prolonged dry spell do not usually produce a large impact on streamflow response since most of it becomes absorbed in the wetland.

Outflow – b) Changing Pattern at the Lower Wetland Course

Fig. 8 shows the difference in flow between Flume A and Flume B (72 m downstream) in 1997 (see Fig. 1c). Larger discharges at Flume B compared with Flume A shows that while the bulk of water comes from the southern part of the wetland, water supplied from the northern section of the wetland augmented high flows. This water source is likely provided by residual snowpatches lying on the north side of the wetland adjacent to a steep slope and by ground ice melt. However, this pattern of water flow was not replicated every year. Earlier snowmelt and a longer thaw period in 1998 caused the flow to be greater from Flume A than Flume B by 2.5 times. This suggests the loss of surface water to seepage as the flow leaves the wetland. The contrasts in flow pattern between these two years demonstate that climatic variability can alter the surface flow.

Outflow – c) Subsurface Flow

Fig. 9 shows that groundwater outflow from the wetland in 1998 is sizeable and increased as the storage capacity was enlarged by ground thaw in the gravel zone. The amount of subsurface discharged from this section of the wetland accounts for approximately 72% of the flow descrepancy between Flumes A and B in 1998. Preliminary field measurements in 1999 suggest that this process is recurrent.

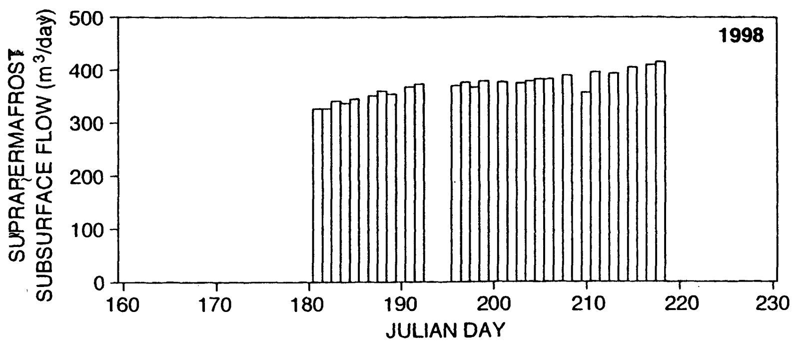

Fig.9. Daily suprapermafrost subsurface flow exiting the wetland between Flumes A and B, 1998.

Storage

Water storage in the wetland can take several forms: within the snow cover, in shallow depressions, within vegetation, peat, and in the mineral soil. Meltwater and precipitation recharge small pools, micro-despressions and saturate the thawed ground. The wetland storage pattern over the three summer seasons shows that the storage remains high and is generally similar every year (Fig.10). In 1996, storage increased over the summer season from 150 to 175 mm, owing to the increased storage capac-

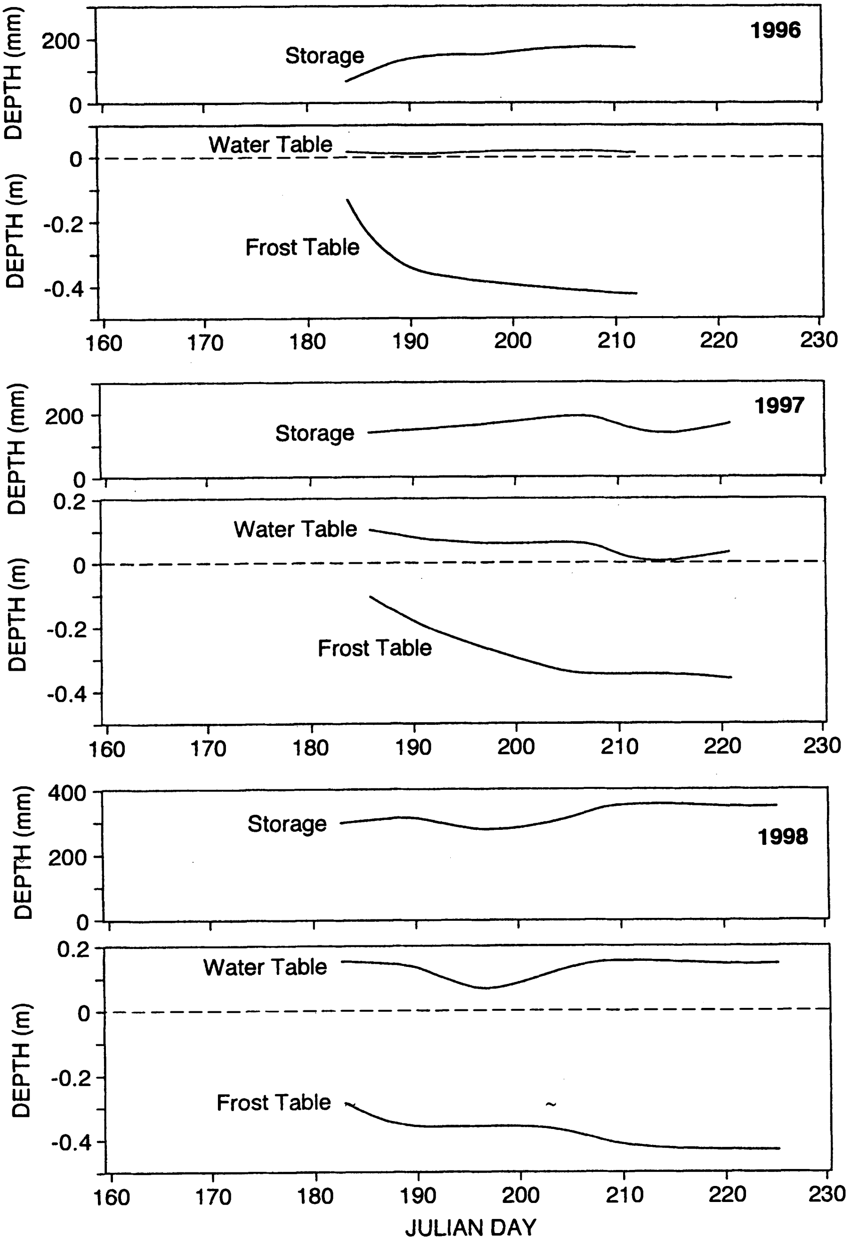

Fig.10. Pattern of storage in the wetland during three summer field seasons (1996-1998). Average storage is plotted in relation to the average frost table development and water table profile for each year.

ity, while the water table remained high. In 1997, the storage was lower than 1996 due to a shallower ground thaw. Slight declines in storage reflect removal of water through evaporation which brings about a drop in the water table. Water storage was higher in 1998 owing to a larger storage capacity and prolonged lateral water inputs. The storage trend reflects the water table fluctuations where a slight drop between Julian Days 190 and 200 can be attributed to elevated evaporation rates (Fig. 6). Large water storage in the wetland is maintained by lateral groundwater inputs from the surrounding areas and reduced evaporation rate in late summer.

Wetland Streamflow Pattern

Church (1974) suggested that flood flows in northern wetlands are greatly attenuated because of the large water retaining capacity of muskeg vegetation, as well as the high resistance to runoff due to the irregular surface. However, Fig. 7 suggests that the seasonal pattern of stream response from the wetland is typical of the nival regime. Snowmelt generates substantial discharges and the steep rising and falling limbs of the hydrographs indicate fast wetland response to water inputs due to a limited storage capacity. The peak discharges are around 0.03 m^3/s and such flows can be sustained by meltwater from the surrounding basin areas. Low flows can be caused by dry spells in 1996 and 1998 where pools become disconnected and can also be brought to a halt by summer snow events (1997). Small rain events help to raise the storage levels in the wetland, but it was the large rain and summer snow events in all three years that generated large streamflow after the main snowmelt season.

Year to year variation in climate brings about fluctuations in the wetland hydrological behaviour, as duration and magnitude of inundation vary according to the wet or dry cycles. Fig. 7 indicates the variability in the timing of snowmelt and runoff, the magnitude and duration of peak flows as well as periods of low to zero flow in this three year period. Climate plays an important role in influencing the mechanisms of the water yields and losses from the wetlands (Woo 1993).

Discussion and Conclusions

1) This patchy high arctic wetland receives water from multiple sources. It is sustained by snow and ground ice melt within the wetland, lateral water inputs from the surrounding basin and summer precipitation. The mechanisms defining this wetland differ slightly from the valley bottom wetland described by Glenn and Woo (1997). The valley-bottom wetland received snowmelt from the surrounding slopes initially, but then it became detached from the slopes and had to rely on summer precipitation and ground ice melt to sustain it. However, in this study, the connections between the wetland and its surrounding landscape are important in defining its hydrological behaviour and sustaining flow throughout the summer seasons.

2) Storage is high at all times in the wetland and is due in part to the multi-sources of water which augment it. Storage is fed by snowmelt, precipitation, lateral water inflows and considerable ground ice melt especially in the early part of the season. In the study of Glenn and Woo (1977), the storage pattern of the valley-bottom wetland declined rapidly after snowmelt due to high evaporation rates and no lateral water inputs.
3) Owing to the frequently saturated status of the wetland, its response to inputs of snowmelt and sizeable rainfall inputs is rapid (see Figs. 2, 3, 7) and the recession limbs are steep reflecting rapid drainage. This pattern of rapid stream response indicates that the wetland system is well connected and inputs of snowmelt and rain can flow through the system with little restriction. However, low intensity rainfall events do not produce noticeable streamflow response, most of the rainwater being absorbed.
4) Given that the there is little flow attenuation, the hydrological response of the wetland is more like a nival regime rather than a muskeg regime. There is no extra capacity to hold water except for small rain or snow events. Climate does influence the mechanisms of the water yields and losses from the wetlands. Variations arising from short-term fluctuations and long-term trends have indirect and direct bearings on these mechanisms, thus affecting the behaviour of the wetland (Woo 1993).
5) Mechanisms controlling the response of this wetland are comparable to other northern wetlands during the snowmelt runoff period when discharges are large (Glenn and Woo 1997; Roulet and Woo 1986; Ryden 1977). As in other studies, evaporation is also an important loss of water from the wetland (Glenn and Woo 1997; Rovansek *et al.* 1996). However, linkages of this wetland with the surrounding area is important in sustaining streamflow through the summer period. Other wetland systems become separated from their surrounding landscape following snowmelt (Glenn and Woo 1997; Rovansek *et al.* 1996) and therefore cannot take advantage of this added water source. This has implications for plant growth and water chemistry.

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