Role of snow in the hydrology of a High Arctic riparian wetland

Kathy L. Young

ABSTRACT

Riparian wetlands are narrow strips of saturated and vegetated ground forming critical links between dry ground and waterways. The hydrology of a riparian wetland situated within a polar oasis landscape near Eastwind Lake, Ellesmere Island, Nunavut (80°80′N, 85°35′W) was investigated in 2006 using a combination of fieldwork and modelling. Supplemental information from 2005 was also employed. This study showed that deep snow in the nearby stream channel does not promote a period of extended over-bank flooding but instead initially serves as a dam blocking most streamwater from entering and flooding the wetland. It was not until the snow dam melts and disintegrates in response to favourable weather conditions that the wetland becomes flooded and fully recharged. This was a delay of three weeks from the previous year. For the remainder of the 2006 growing season, contributions of meltwater from late-lying snow beds located within and adjacent to the stream channel and near the headwaters were essential for maintaining saturated conditions in the wetland.

Key words | Arctic hydrology, channel snow, High Arctic, permafrost, riparian wetland

Kathy L. Young
Department of Geography,
York University,
4700 Keele Street,
Toronto, Ontario M3J 1P3,
Canada
E-mail: *klyoung@yorku.ca*

INTRODUCTION

Riparian wetlands are unique strips of saturated and vegetated ground forming important links between terrestrial landscapes and aquatic zones. They serve to both modify and be modified by fluvial and chemical processes and have been well studied in temperate environments (e.g. Toner & Keddy 1997; Hauer & Smith 1998; Cole & Brooks 2000; Vidon & Hill 2006). Cole & Brooks (2000) suggest that these wetlands are the wettest when compared to other wetland types and, with some exceptions, show the smallest range in hydrologic behaviour. They indicate that duration of inundation and saturation for most riparian sites is about 81%. Toner & Keddy (1997) suggest that the duration of flooding and the frequency of flooding is important for determining plant type structure. Woody plants succeed herbaceous plants in areas with both infrequent flooding and duration. Vidon & Hill (2006) and others have focused on defining the biogeochemistry of these zones and their ability to deplete nitrogen-rich waters draining from upslope agricultural fields.

These narrow wetlands are also common features in High Arctic landscapes running along streams and rivers, yet their hydrology is not well understood. Woo & Young (2003) provide some information on their hydrology through their study on Cornwallis Island, Nunavut, an environment considered to possess a polar desert regime. These areas generally have cool, wet springs and summers, and normally receive more snow than polar oasis landscapes (Young & Woo 2004*b*). Woo & Young (2003) found that water tables in the riparian wetland on Cornwallis Island continually remained high from seasonal snowmelt runoff and extended over-bank flooding from snow-choked stream channels. Due to diluted conditions, electrical conductivity and water cation levels remain low, in comparison to groundwater-fed depression-type wetlands.

Here, I describe the hydrology of a riparian wetland situated within a polar oasis landscape near Eastwind Lake, Ellesmere Island, Nunavut (80°80′N, 85°35′W) during the

doi: 10.2166/nh.2008.004

2006 field season. Information from the 2005 field season is also used. Unlike the Woo & Young (2003) study, snow in the channel does not promote a period of extended over-bank flooding but in fact initially serves as a dam, blocking streamwater from entering and flooding the wetland prior to its disintegration. This study investigates how the wetland responds to these conditions along with contributions of meltwater from late-lying snow beds which exist in the stream channel and along steep slopes and valleys near the headwaters. A combination of field data (climate, hydrology) and a snowmelt model (Woo & Young 2004) are employed to explore the dual role of snow (blockage versus recharge) in the hydrology of a High Arctic riparian wetland.

STUDY AREA

This study took place from early May to early August 2006 near Eastwind Lake, Ellesmere Island, Nunavut (80°80′N, 85°35′W). Additional information from the 2005 field season at this site supplements this investigation. The wetland is situated about 20 km inland from Eureka, a government weather station (Figure 1). The wetland site is composed of a broad floodplain characterised by a main stream channel running through it and a series of water pathways or streamlets. Similar riparian wetlands occur in the area and are likewise associated with creeks draining steep hills which border and confine the larger Eastwind Lake lowland (Nettleship & Smith 1975). Elevation within the studied wetland ranges from 142 m at its inlet to 136 m a.s.l. at its outlet: a gradient of 0.05. Parts of the wetland are well vegetated and contain much wet meadow-type vegetation (e.g. cotton-grass, moss, sedges and grasses), while other zones are void of plant life and are better described as gravelly mudflats. These latter areas are routinely subjected to stream waters which deposit much silt and detritus here, as the current slows through the area. A soil profile dug in the middle of the wetland indicates a thin, organic layer over a saturated, gravelly–silty soil grading to clay with depth. Orange iron stains indicative of water-logged conditions occur throughout the soil profile. The area is below the marine limit (ca. 150 m a.s.l, 8.7 ka BP) (Bell 1996). The stream which floods and recharges this wetland originates near the peak (ca. 590 m) of Black Top Ridge, a ridge which runs in a southwest to northeast direction across the Fosheim Peninsula, from Eureka Sound to Greely Fiord. This stream, referred to here as Black Top Ridge Creek (BTRC), is just one of many (>20) which drain this high ridge and empty into adjacent wetland zones. Snow typically persists on the ridge and in steep gullies much longer than the lowland wetland, owing to the cooler conditions here.

The study area can be described as having a polar-oasis-type climate (Woo & Young 1997; Woo & Guan 2006). It typically experiences warmer and drier conditions than elsewhere in the High Arctic, as it is sheltered by nearby mountains from low pressure systems that originate in the Arctic Ocean (Maxwell, 1980; Young & Woo, 2004*a*,*b*; Woo & Guan 2006). Hence, snowfall and rainfall are lower than polar desert regimes (Woo & Young 1997). Moreover, due to its interior position, summers at the wetland are also warmer than the coastal Eureka station (see Woo & Guan 2006).

Upon arrival at the field site in early May 2006, the snow cover was thin over the uplands and was rapidly being consumed by sublimation (Woo & Young 1997). The bulk of the winter snow pack had been blown into low-lying areas (i.e. stream channels and depressions). However, due to a lingering cool, cloudy episode, the main snowmelt season was delayed until 24 May and persisted for 21 d, 11 d longer than in 2005. Overall, the 2006 season was much cooler (ca. 400 thawing degree-days) and wetter than 2005, which was one of the warmest on record for this area (over 500 thawing degree-days was recorded at Eureka) (see Woo & Guan 2006). Summer rainfall in 2006 amounted to 39 mm up until 8 August (end of fieldwork) and then an additional 30 mm fell between 9–14 August at Eureka (Woo & Guan 2006). Only 27 mm fell in 2005 at the study site, in comparison to 19 mm at the Eureka weather station. Growing conditions were less favourable in 2006 than 2005. In 2006, growing-degree days totalled 138 versus 224 in 2005. This resulted from 17 fewer growing days (>5°C) in 2006 than 2005, when a similar time span is considered.

METHODOLOGY

Field

Five transects were laid out across the riparian wetland and adjoining stream channel and a snow survey was

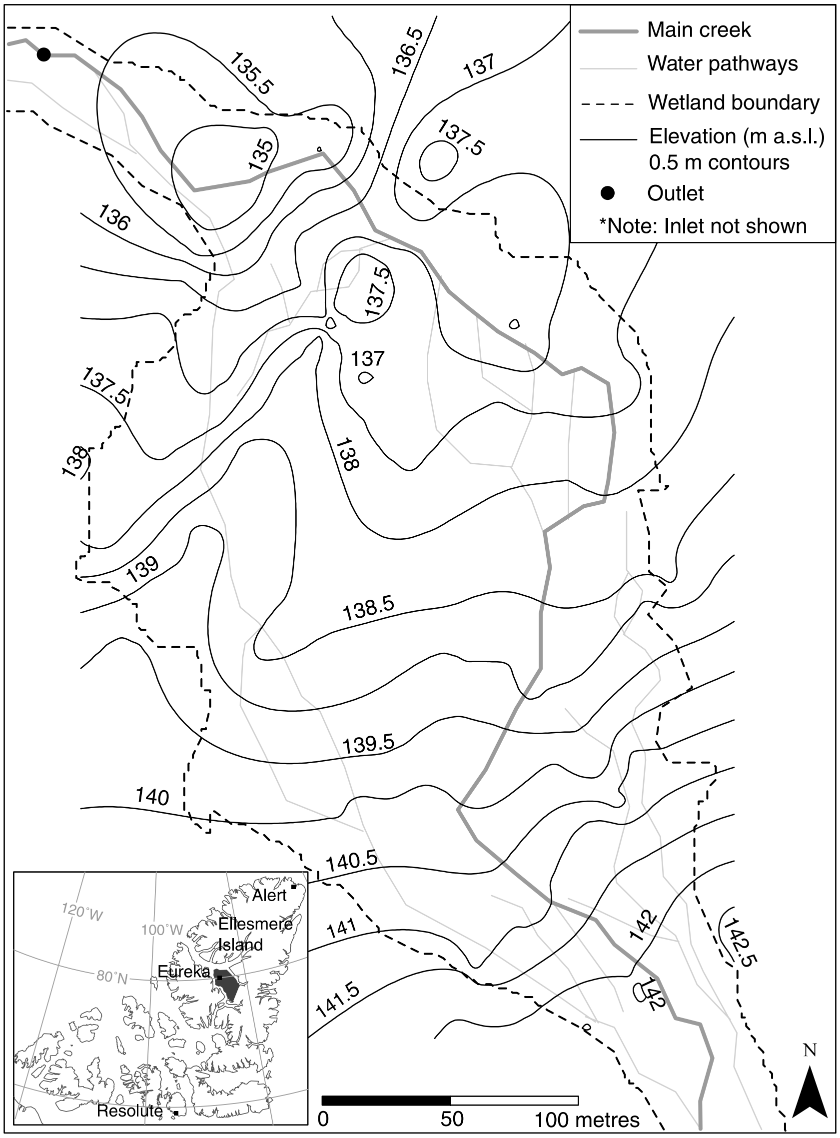

Figure 1 | Topographic map of the riparian wetland (80°80′N, 85°35′W) located 20 km north of Eureka, Ellesmere Island, Nunavut, Canada (see inset map). The Fosheim peninsula is shaded.

conducted, following after Woo (1998). Snow depth measurements were made every 10 m and snow density was determined at the beginning and end of each transect with an Environment Canada MSC snow tube. Direct measurements of snowmelt followed after Heron & Woo (1978). Snow ablation lines were installed to measure daily lowering of the snow surface, which was converted into water equivalent units using the surface snow density.

In 2005, three permanent transects were established across the wetland (see Figure 2) and a series of perforated and screened water wells (3–4) per transect were installed down to the permafrost table. In the post-snowmelt period (2006), daily water levels were measured at the wells with electronic water sensors, and depth of thaw was probed weekly at 10–12 locations along each transect (see Figure 2). The saturation state of the wetland was classified into three classes and followed the approach described by Cole & Brooks (2000). Here a depth of -15 cm below the ground surface was set as the limit of the rooting zone. Inundation was defined when water tables (WT) were >0 cm; saturation (WT $= 0$ to -15 cm); and dry conditions (WT < -15 cm). Near-surface soil moisture (upper 5 cm) was determined at

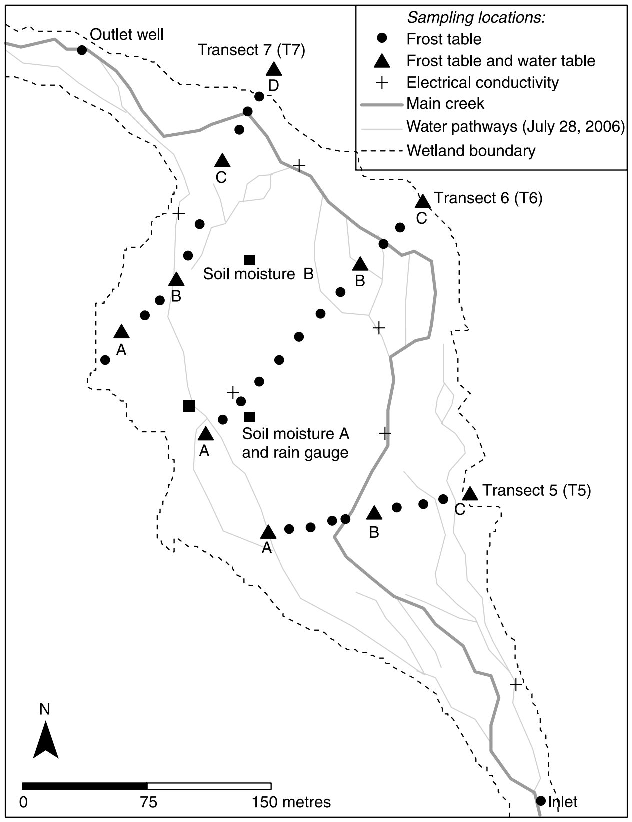

Figure 2 | Series of transects across the wetland indicating water wells and frost table locations. Stream gauging locations at the inlet and outlet are indicated, as are the rain gauge, soil moisture and electrical conductivity sampling sites. Note that electrical conductivity measurements were also made at the stream's inlet and outlet.

two locations in the middle of the wetland (Figure 2) using the gravimetric approach with values later converted into volumetric units (Woo & Guan 2006). These soil moisture measurements were made in association with others (e.g. wet meadow, tundra upland, mesic ground, pond rim) as part of a broader wetland study.

An automatic weather station (elevation = 138 m) situated over a wet meadow provided hourly meteorological information: net radiation (W/m^2), incoming and outgoing solar radiation (W/m^2), air temperature (°C), relative humidity (%) and wind speed (m/s). Summer precipitation (mm) was measured with a recording tipping-bucket rain gauge and verified with four manual rain gauges, one of them situated in the middle of the riparian wetland. A stilling well was situated near the outlet of the riparian wetland in the stream channel and an Ecotone water level recorder measured the stage here (±10 mm). Stage levels were corrected routinely with direct depth measurements. Current metering occurred at both the inlet and outlet locations (see Figure 1) usually two to three times per day during high flows and once per day in low flow conditions. This allowed rating curves to be determined for each site: inlet location $Q = 4.98H^{2.23}$, $r^2 = 0.93$, $n = 38$ and for the outlet $Q = 4.98H^{2.14}$, $r^2 = 0.94$, $n = 38$. Here, Q is stream discharge in L/s; H is the stage in the barrel (cm) and n is sample size. A continuous record of reliable discharge was determined for both locations from 29 June onwards. Electrical conductivity was measured with a WTW Tetra conductivity probe ($1 \pm \mu S/cm$). The stream (inlet, outlet) and areas across the wetland ($n = 5$) were sampled daily before 1200 h EST (see Figure 2). A topographic survey of the study site occurred in late July using a transit level and stadia rod. Elevations were tied to a known benchmark.

Snowmelt model

For this study I wanted to identify and understand the processes which were driving the stream flow pattern passing through this riparian wetland. An initial comparison of stream discharge to both air temperature and net radiation proved inadequate. A physically based snowmelt model (Woo & Young 2004) was then employed to assess the processes (energy receipt vs. rain input) modifying stream flow. The model has the capability of being able to reconstruct the snow pack in a variety of terrain zones or units (e.g. flat areas, valleys, plateaus and slopes of various aspects and angles) in relation to a "base station" and then model daily melt according to the surface energy approach. The utility of this model has recently been confirmed at another wetland site on Somerset Island, Nunavut (Young & Abnizova 2005).

Inputs to the model include both climate and snow information from the base station, and outputs consist of daily energy receipt and snowmelt (see Appendix 1 for model structure). Albedo decay in the model follows the empirical function of Woo & Dubreuil (1985) with an adjustment made for lingering snow packs (e.g. late-lying snow beds). Within the model the cold content of the snow pack at each site is adjusted on a daily basis and ablation of the snow pack (snow water equivalent (SWE), mm) only occurs once the cold content is eliminated. The model has the ability to adjust for lower wind speeds in valleys and can modify incoming solar radiation for sloping terrain. Long-wave radiation outgoing is modelled using $\varepsilon_s = 0.97$, modelled surface temperature (T_s) and incoming long-wave radiation is determined by air temperature and ε_a, the emissivity from the sky according to Idso & Jackson (1969) using air temperature (°C). Radiation melt attributed to net radiation is composed of two components: the short-wave part calculated using snow albedo (α) and the long-wave part calculated as the balance between incoming and outgoing long-wave radiation. The bulk transfer approach (Price & Dunne 1976) is used to estimate the turbulent fluxes, with temperatures adjusted for elevation differences (i.e. base station versus the terrain unit). Hourly climate data for the snowmelt model came from the automatic weather station and includes incoming solar radiation (W/m^2), air temperature (°C), relative humidity (%), wind speed (m/s) and precipitation (mm) except for atmospheric pressure (Pa) which was obtained from the Eureka weather station, 20 km to the south.

Initial snow information was limited to the riparian wetland (base station). Given that most stream flow was likely generated from snow further upstream, snow information for this valley zone was required. Using a series of 2005 photos obtained from students hiking up the creek channel to the top of Black Top Ridge on 1 July, I dissected the stream channel into a series of sections based on slope, aspect, elevation and snow amount (i.e. I compared snow conditions in the photo to the wetland snow cover to derive an estimated snow index for the terrain) (see Table 1). A snow index is the ratio of the snow pack at the terrain unit, in this case, a valley in comparison to the snow amount at the base station. Normally, extensive snow surveys should have been conducted along the stream channel in 2006 but this was not logistically possible (e.g. no snowmobile and treacherous terrain). However, reliable snow pack information at the base station is required to establish realistic snow conditions at modelled sites. Initially, the mean snow survey at the riparian site, which included the stream channel, yielded an estimate of 106 ± 50 mm. Review of the records indicated that this initial snow survey was biased towards the flat wetland zone, where shallow snow cover existed rather than the stream channel where deeper snow had been trapped, and which likely provided a better reflection of snow conditions further up the stream channel. Moreover, our snow records indicated that on occasion we were unable to measure deep snow in the wetland stream channel due to limitations of our snow rod and snow density tube. This would also serve to underestimate the wetland snow pack (base station).

Snow pits dug in the stream channel might have confirmed its "true" snow depth and density but due to time constraints during the snow survey, which included 12 other sites, and rapid sublimation, it was not possible to make these detailed snow-pit measurements. For modelling purposes, a decision was then made to adjust the initial snow amount in the wetland upwards by 50% (159 mm snow water equivalent) to ensure that a more realistic "base station" snow cover existed. This amount was comparable to the mean snow water equivalent derived from two transects in the wetland which crossed the main stream channel (e.g. TR1 = 172.9 mm and TR2 = 147 mm), an area which had more snow and was perhaps more indicative of the snow conditions in the stream channel. Finally, this adjustment to snow water equivalent at the base station allowed valley snow indices for the modelled sites to be held to reasonable levels based on previous surveys of snow-filled valleys across Cornwallis Island (Woo & Young 2004).

Table 1 | Initial conditions for the snowmelt model (see Woo & Young 2004)

Terrain	Snow index	Mean elevation (m)	Aspect (deg)	Slope angle (deg)	Area (m^2)
Base (swe = 159 mm)	1.00	138	–	–	
Valley 1	1.00	134	315	0.01	97,656
Valley 2	3.00	160	315	4	97,656
Valley 3	4.5	220	315	19.7	317,383
Valley 4	6.0	350	315	19.7	195,312
Valley 5	6.0	500	315	19.7	67,139

Photographs taken during both 2005 and 2006 also indicated that most snow was constrained to the main stream valley and nearby slopes. This information served to define areas for the different contributing zones (m^2). Here, a 1:50,000 topographic map of the Black Top Ridge area (Series A701, Map 340 B/3, Edition 1) was employed. Modelled stream flow was generated from both simulated melt and measured rainfall inputs obtained from the weather station. These data were areally weighted for each terrain unit. Table 1 provides the site conditions for the base and the series of valley terrain units comprising the stream channel. Isothermal conditions were assumed for the base station since the seasonal snow pack had largely disappeared from the lowland and only snow in the stream channel and valleys remained.

RESULTS AND DISCUSSION

Model results

Modelled and measured daily stream flow is found in Figure 3. The stream flow pattern, shared by both curves, shows two distinct periods: an initial peak discharge followed by rapid drainage which often occurs when a valley snow dam releases ponded water (Woo & Sauriol 1980) and then a second period of moderate and sustained flow, punctuated by pulses of lower discharge, a rhythm often replicated by melting late-lying snow beds in response to changing weather conditions (e.g. Marsh & Woo 1981; Lewkowicz & Young 1990). A reasonable relationship between measured and modelled daily stream flow (Figure 3) at the outlet (within 5%) provides confidence in the initial conditions selected and the types of processes controlling stream flow through the wetland (see below). Differences between model and measured results can be attributed to lags in the system, e.g. meltwater being stored for several days in the valley snow pack upstream of the basin inlet, a snow dam in the channel delaying flow (Woo & Sauriol 1980), situations which cannot be reproduced by the model. Both over- and under-estimates of measured stream flow also arise from errors in assessing initial snow amounts and areal coverage. Overall, this adequate performance is to be expected, given the lack of snow information for this mountain stream.

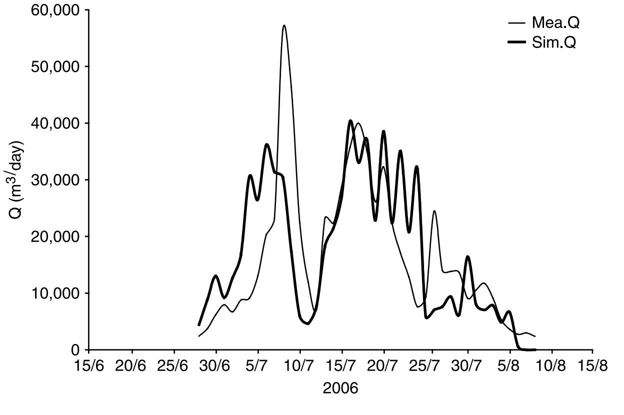

Figure 3 | Simulated and measured stream flow, Black Top Ridge Creek (BTRC), 2006.

Snow-dam period

Figure 4(a) indicates the saturation conditions of the wetland prior to the large flood event of 9 July (Figure 3). During this period the saturation levels are reduced, with water tables falling below the ground surface (Figure 5) and

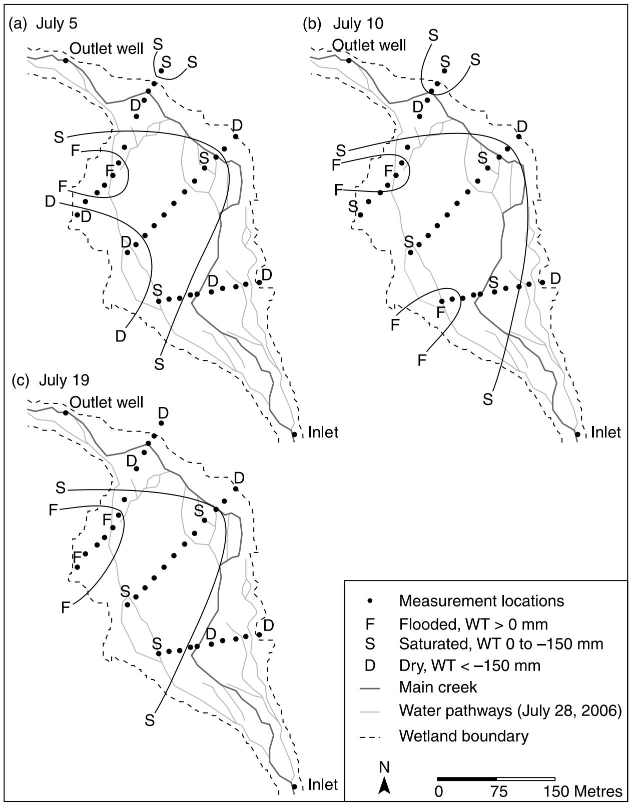

Figure 4 | Water saturation patterns in the riparian wetland, 2006.

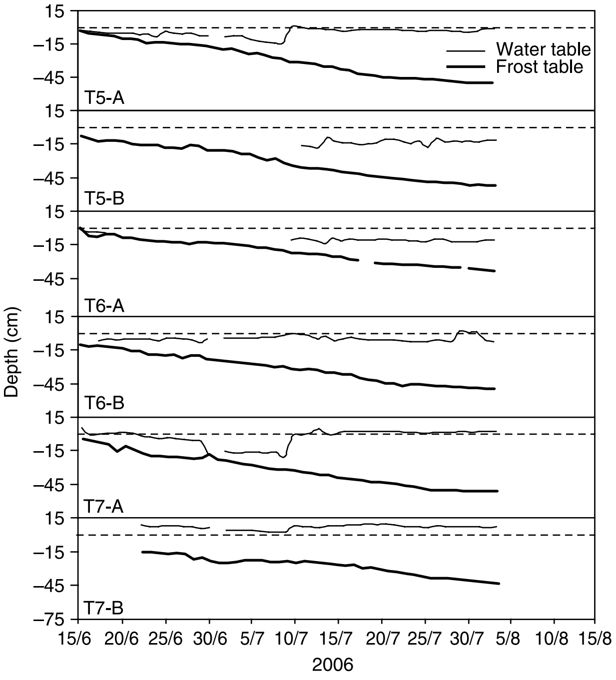

Figure 5 | Seasonal water table fluctuations and frost table pattern for selected well locations along wetland transects.

the occurrence of dry water wells (not shown). Soil moisture levels while limited in extent confirm this drying pattern (i.e. $\theta_s < 100\%$ vol. water content). Electrical conductivity values are also variable, especially for the wetland zone which had much higher and erratic values than channel water and minor flow paths (Figure 6). Dall'O *et al.* (2001) indicate that dry intervals are common for riparian wetlands and are not considered critical, e.g. these episodes help to aerate the soil, yet Toner & Keddy (1997) indicate that saturated conditions are required for germination and infrequent flooding can lead to a change in plant structure. They discovered that, for temperate wetlands, the duration of floods and the period between initial and secondary floods were key factors in preventing woody substrates from succeeding herbaceous species. Drier conditions in the Athabaska Delta after a water diversion led to a shift in vegetation from herbaceous to woody plants (Toner & Keddy 1997).

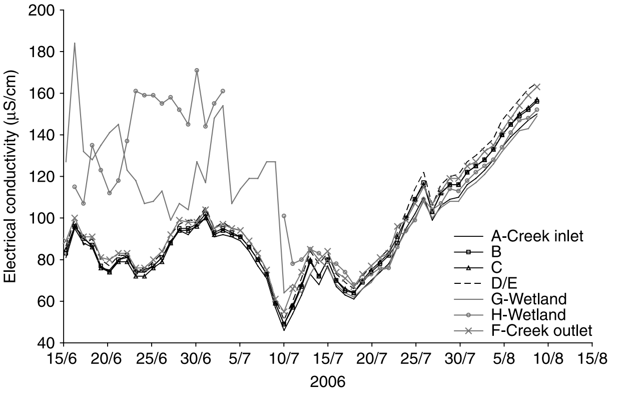

Figure 6 | Seasonal pattern of electrical conductivity at selected locations in the wetland and stream channel.

Flood period

A significant discharge event (see Figure 7) did not occur until a snow dam diverting most of the seasonal snowmelt flow (A) from the riparian wetland was finally destroyed on 9 July (labelled here as B). Model results indicate that this large release of water was triggered by a stretch of sunny, warm and windy conditions (enhanced net radiation and sensible heat flux receipt) which accelerated melting after an unusually long and cool melt season. This pronounced flood event allowed water tables to rise (Figure 5) often above the ground surface, expanded the saturated zone (Figure 4(b)) and diluted the wetland, as noted by the large drop in electrical conductivity (Figure 6). This event, which rapidly recharged the wetland, happened about three weeks later than in 2005. The occurrence of snow dams and snow-choked channels and their ability to delay and impound water levels is common in High Arctic environments (Woo & Sauriol 1980; Xia & Woo 1992). During the winter period, snow infills the dry valleys and drifting reshapes the snow

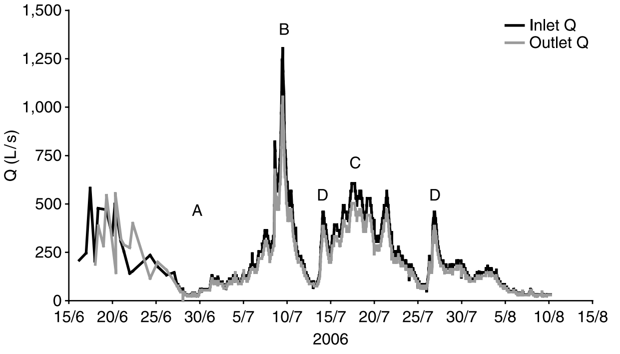

Figure 7 | Seasonal stream flow pattern of Black Top Ridge Creek (BTRC), 2006. Here A refers to a low stream flow period, B indicates the flood event from the release of a snow dam, C indicates melt from late-lying snow beds and D indicates stream flow peaks driven by rain and melt generated from rain on snow.

mass into ridges and troughs which, during the initial runoff period, become dams and pools as meltwater enters the valleys. Until the snow dams are breached either by seepage, tunnelling or water spillage over the dam there is no integrated flow. Hydrologically, the bursting of snow drift dams causes a sudden release of water, thus accelerating the break-up of the snow-choked channel below the dam (Woo & Sauriol 1980). Once the flow network is established, the stream has to cut vertically and laterally through the residual snow. If the snow in the valley is deep, undercutting and collapse of snow blocks can also occur, sometimes blocking stream flow for short periods of time until the snowblock disintegrates or the stream finds a new route.

Late-lying snowmelt period

Figure 7 indicates that discharge levels were steady throughout July and did not drop off until early August when cooler and cloudy conditions became more frequent. The snowmelt model suggests that this period of enhanced flow (indicated by C) arises from the melting of channel snow beds and lingering snow on steep slopes adjacent to the channel. The pulses or sharp increases are induced by sunny weather along with warm and windy conditions which enhance fluxes of sensible heat. Young & Lewkowicz (1990) similarly found that net radiation and sensible heat dominated melt at a large perennial snow bank, near Ross Point, Melville Island, where the highest discharges occurred on clear, warm and windy days. Stream flow peaks labelled as D are driven by rain and the melt generated by rain on snow. Energy levels were low (cloudy, cool and calm) for these days but rain helped to elevate the importance of the precipitation heat flux and its ability to melt snow. These residual snow beds and the meltwater that they produced were important for keeping the wetland saturated for the remainder of the summer (Figure 4(c)). Figure 5 reveals that most water tables remain elevated for the remainder of the growing season. The persistence of elevated soil moisture values (i.e. $\theta_s = 100\%$ vol. water content) similarly confirms this pattern. Electrical conductivity values start rising during this post-peak period but increases are similar amongst sites suggesting stable moisture conditions throughout the wetland (Figure 6).

Comparable observations at this site in 2005 suggest that this period of "secondary flooding" after the main melt season is a regular event. Without these additional water inputs it is doubtful whether the wetland would have remained saturated given the potentially high losses of evaporation which can occur from wetland surfaces. For example, in 2005, Woo & Guan (2006) found that evaporation exceeded 4 mm/d for about 10 d here and averaged 3 mm/d for 60 d.

CONCLUSIONS

For this riparian wetland which is found in a polar-oasis-type climate, snow plays a leading role in controlling its hydrology. First, the snow-choked channel initially deprives the wetland of much stream water, a delay of three weeks over the previous year. Delay in the arrival of these meltwater results in the shrinkage of the saturated zone as the water table falls below the ground, often below the rooting zone, altering its water chemistry. Disintegration of a valley snow dam allows floodwater and meltwater from residual channel snow to enter the wetland, promote inundation and expand the saturated zone. Electrical conductivity levels drop in response to diluted conditions.

Maintenance of wetland saturation for the remainder of the growing season can be attributed to the melting of late-lying snow beds either in the stream channel itself or on nearby steep channel slopes. Net radiation and sensible heat flux were the main drivers behind this melt, while rain was important in triggering stream flow through its influence on the melt flux.

This was a one-time-only study and its applicability to other riparian wetlands in the vicinity is limited. Due to a helicopter crash at our field station in 2006, reconnaissance of similar ecosystems in 2006 was not possible. However, an important finding of the study is that more attention should focus on channel snow and the role of this snow in steep catchments. Snow storage in these zones is important in providing meltwater to low-lying areas long after the seasonal snow pack has disappeared. Our study showed that this meltwater was essential for keeping a riparian wetland saturated for most of the summer. Loss of this snow and its ability to recharge these types of wetlands on a regular basis might lead to drier conditions and eventually a shift in vegetation type (e.g. sedges to *Salix arctica*). Ultimately, this could impact the grazing patterns of local animals, especially caribou and muskox.

ACKNOWLEDGEMENTS

The author would like to acknowledge the Natural Sciences and Engineering Research Council of Canada for funding, along with financial assistance from the Northern Student Training Program and York University. Thanks are due to the Polar Continental Shelf Project for excellent logistical support and to May Guan and Kara Pul for their hard work and dedication in the field. May Guan also supplied photos from the 2005 field season and helped to clarify questions during the write-up of the study. Laura Brown's assistance in the production of maps and diagrams is gratefully acknowledged. I am grateful for the constructive comments of two anonymous reviewers and those of Professor Lars Bengtsson.

REFERENCES

Bell, T. 1996 The last glaciation and sea level history of Fosheim Peninsula, Ellesmere Island, Canadian High Arctic. *Can. J. Earth Sci.* **33**, 1075–1086.

Cole, C. A. & Brooks, R. P. 2000 Patterns of wetland hydrology in the Ridge and Valley Province, Pennsylvania, USA. *Wetlands* **20**, 438–447.

Dall'O, M., Kluge, W. & Bartels, F. 2001 FEUWAnet: a multi-box water level and lateral exchange model for riparian wetlands. *J. Hydrol.* **250**, 40–62.

Hauer, F. R. & Smith, R. D. 1998 The hydrogeomorphic approach to functional assessment of riparian wetlands: evaluating impacts and mitigation on river floodplains in the USA. *Fresh. Biol.* **40**, 517–530.

Heron, R. & Woo, M. K. 1978 Snowmelt computations for a High Arctic site. *Proc. 35th Eastern Snow Conference*, Hanover, NH, pp. 162–172.

Idso, S. B. & Jackson, R. D. 1969 Thermal radiation from the atmosphere. *J. Geophys. Res.* **74**, 5397–5403.

Lewkowicz, A. G. & Young, K. L. 1990 Hydrology of a perennial snowbank in the continuous permafrost zone, Melville Island, Canada. *Geogr. Ann.* **72a**, 13–21.

Marsh, P. & Woo, M. K. 1981 Snowmelt, glacier melt and high Arctic stream flow regimes. *Can. J. Earth Sci.* **18**, 1380–1384.

Maxwell, J. B. 1980 Climatic regions of the Canadian Arctic Islands. *Arctic* **34**, 225–240.

Nettleship, D. N. & Smith, P. A. (eds) 1975 *Ecological Sites in Northern Canada*. Canadian Committee for the International Biological Program Conservation Terrestrial–Panel 9, Canadian Wildlife Service, Ottawa, Ontario.

Price, A. G. & Dunne, T. 1976 Energy balance computations of snowmelt in a Subarctic area. *Water Resour. Res.* **12**, 686–694.

Toner, M. & Keddy, P. 1997 River hydrology and riparian wetlands: A predictive model for ecological assembly. *Ecol. Appl.* **7**, 236–246.

Vidon, P. & Hill, A. R. 2006 A landscape based approach to estimate riparian hydrological and nitrate removal functions. *J. AWRA* **42**, 1099–1112.

Woo, M. K. 1998 Arctic snow cover information for hydrological investigations at various scales. *Nordic Hydrol.* **29**, 245–266.

Woo, M. K. & Dubreuil, M.-A. 1985 Empirical relationship between dust content and Arctic snow albedo. *Cold Regions Sci. Technol.* **10**, 125–132.

Woo, M. K. & Guan, X. J. 2006 Hydrological connectivity and seasonal storage change of tundra ponds in a polar oasis environment, Canadian High Arctic. *Permafrost Periglacial Process.* **17**, 309–323.

Woo, M. K. & Sauriol, J. 1980 Channel development in snow-filled valleys, Resolute, N.W.T., Canada. *Geogr. Ann.* **62A**, 37–56.

Woo, M. K. & Young, K. L. 1997 Hydrology of a small drainage basin with polar oasis environment, Fosheim Peninsula, Ellesmere Island, Canada. *Permaforst Periglacial Process.* **8**, 257–277.

Woo, M. K. & Young, K. L. 2003 Hydrogeomorphology of patchy wetlands in the High Arctic, polar desert environment. *Wetlands* **23**, 291–309.

Woo, M. K. & Young, K. L. 2004 Modeling arctic snow distribution and melt at the 1-km grid scale. *Nordic Hydrol.* **35**, 295–307.

Xia, Z. J. & Woo, M. K. 1992 Theoretical analysis of snow dam decay. *J. Glaciol.* **38**, 191–199.

Young, K. L. & Lewkowicz, A. 1990 Surface energy balance of a perennial snow bank, Melville Is., N.W.T. Canada. *Arctic Alpine Res.* **22**, 290–301.

Young, K. L. & Woo, M. K. 2004*a Queen Elizabeth Islands: problems associated with water balance research*. IAHS Publ. 290, 237–248.

Young, K. L. & Woo, M. K. 2004*b Queen Elizabeth Islands: water balance investigations*. IAHS Publ. 290, 152–163.

Young, K. L. & Abnizova, A. 2005 High Arctic ponds, Somerset Island, Nunavut: Spatial and temporal variations in snowcover and snowmelt. *Proc. 62nd Eastern Snow Conference*, Waterloo, Ontario, June 7–10, pp. 69–82.

First received 2 November 2007; accepted in revised form 8 March 2008

APPENDIX 1

Flow chart of surface energy balance snowmelt model (after Woo & Young 2004)