ABSTRACT

The seasonal snowcover and snowmelt (2008–2010) of an extensive low-gradient wetland at Polar Bear Pass, Bathurst Island, Nunavut, Canada (75°40’ N, 98°30’ W) was examined. This wildlife sanctuary is characterized by two large lakes and numerous tundra ponds, and is bordered by rolling hills with incised hillslope stream valleys. In arctic environments snow remains one of the most important sources of water for wetlands. End-of-winter snowcover measurements (snow depth, density, water equivalent) together with direct and modeled estimates of snowmelt provided an assessment of the seasonal snowcover regime of representative terrain types comprising upland (plateau, stream valley, late-lying snowbed) and lowland landscapes (wet meadow, ponds, lakes). In all three seasons, deep and persistent snowpacks occurred in sheltered areas (stream valleys) and in the lee of slopes adjacent to the wetland. Exposed areas yielded shallow snowpacks (e.g. plateau, pond) and they melted out rapidly in response to favorable weather conditions. Overall, the basin snowcover and melt progression was dominated by accumulation and melt occurring in upland areas. We surmise the sustainability of this low-gradient wetland is dependent on snowmelt contributions from upland sites.

Key words: arctic snowcover, arctic wetland, snowmelt, uplands

INTRODUCTION

Wetlands are critical landscapes in the High Arctic, providing food for migratory birds and larger fauna, such as caribou and muskox, and storing and replenishing freshwater supplies. They have recently been the focus of interest in terms of their role in up-taking and releasing greenhouse gases such as carbon dioxide (CO₂), methane (CH₄) and water vapour (H₂O). Snow remains an important source of water for these ecosystems, often replenishing ponds and lakes and re-saturating wet meadow areas at the end of a cold winter season. Since the mid-1990s, we have investigated the hydrologic processes, including snowcover distribution and snowmelt, in a range of wetlands from the small patchy type (local scale, 1–10 km²) up to the meso-scale (c. 25 km²).

We have a reasonable understanding of the hydrologic dynamics of these ecosystems and their water budgets, as do others (e.g. Woo & Guan 2006; Woo et al. 2006; Woo & Young 2006; Young & Abnizova 2011). These studies have confirmed that, except for exposed plateau areas, most upland areas such as slopes and deep stream valleys can collect much more snow than wetland areas (Table 1). During spring melt, these upland areas can provide a large influx of surface water that saturates adjacent wet meadow zones and fills ponds. This is a critical process in High Arctic wetland environments, where summer rainfall can often be insufficient to overcome water losses to evaporation or vertical seepage into thawing ground. The understanding of linkages between upland and lowland areas must also be taken into account when considering future climate warming in cryosphere landscapes (ACIA 2005; Brown et al. 2007). Possible shorter snow seasons will decrease the albedo, thereby increasing radiation absorption and further enhance the warming of the surface atmosphere (Déry & Brown 2007; Euskirchen et al. 2007). Woo & Guan (2006)
suggest that future warmer temperatures will modify snowmelt and enhance ground thaw and evaporation losses in wetlands, but they will also affect adjacent upland areas. They hypothesize that water contributions from upland areas to low-lying wetlands could be compromised.

In 2008, our hydrologic studies shifted from small to much larger wetland systems existing in the Canadian High Arctic. In the 1970s and 1980s, the ecology of these large wetland sites had been extensively studied (Nettleship & Smith 1977). Little was known of their hydrology and role in the broader landscape, however, such as control on runoff into adjacent polar seas, uptake/loss of greenhouse gases or terrestrial carbon movement (Abnizova et al. 2012). To date, only one study has examined the snowcover regime of an extensive wetland system (Truelove Lowland; Rydén 1977), yet here the focus was only on a small wet meadow area (0.12 km² intensive watershed).

The objectives of this study are to: (1) present current understanding of the snowcover characteristics (depth, density, snow water equivalent (SWE)) and snowmelt processes in response to climatic conditions, over a 3-year period at Polar Bear Pass (PBP), an extensive low-gradient wetland centrally located on Bathurst Island; and (2) assess upland/lowland snowcover and melt patterns (e.g. timing, duration, end-date). We surmise that the sustainability of this low-lying, low-gradient wetland is dependent on upslope snowmelt contributions, as has been observed in other arctic studies (Woo & Guan 2006; Abnizova & Young 2010).

Table 1  |  End-of-winter snow water equivalent (SWE, mm) for various upland/lowland terrain found in the Canadian High Arctic

<table>
<thead>
<tr>
<th>Study location</th>
<th>Plateau (upland)</th>
<th>Slope (upland)</th>
<th>Stream valley (upland)</th>
<th>Wet meadow (lowland)</th>
<th>Ponds (lowland)</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>McMaster River Basin, Cornwallis Island, NU 74 45’ N, 94 50’ W</td>
<td>1-100</td>
<td>101-200</td>
<td>&gt;400</td>
<td>N/A</td>
<td>N/A</td>
<td>Woo (1983)</td>
</tr>
<tr>
<td>Creswell Bay, Somerset Island, NU 72 43’ N, 94 15’ W</td>
<td>153-258 (upland ponds)</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
<td>148-163</td>
<td>Abnizova &amp; Young (2010)</td>
</tr>
<tr>
<td>Fosheim Peninsula, Ellesmere Island, NU 79 58’ N, 84 28’ W</td>
<td>44</td>
<td>117</td>
<td>164</td>
<td>N/A</td>
<td>N/A</td>
<td>Glenn &amp; Woo (1997)</td>
</tr>
<tr>
<td>Eastwind Lake, Fosheim Peninsula, Ellesmere Island, NU 80 08’ N, 85 35’ W</td>
<td>N/A</td>
<td>79-163</td>
<td>N/A</td>
<td>50-95</td>
<td>49-68</td>
<td>Woo &amp; Guan (2006)</td>
</tr>
</tbody>
</table>

**STUDY AREA**

The study took place over a 3-year period (2008–2010) at PBP, which is located in the middle of Bathurst Island (75 40’ N, 98 30’ W; Figure 1). PBP is a designated wildlife sanctuary and is classified as a Ramsar site of international significance. As is typical of wetland complexes, it comprises wet meadows, dry and wet ground, two large lakes and a mosaic of ponds (c. 4,800 ponds, pers. comm. Sina Muster, July 2011) situated on a range of substrate types (coarse to fine-grained), and has a low gradient (c. 0.002, <0.1% slope). Recent remote sensing analysis (pers. comm. Sina Muster, April 2012) together with a geographic information system (GIS) reveals that its entire watershed is 422 km² in area. The wetland runs east–west and is bordered by low-lying hills to the north and south with an elevation rising from c. 23 m up to 150 m a.s.l. The slopes themselves are dissected by a series of v-notched valleys, about 50 of varying order (from first- to fourth-order). They are effective in transferring water and nutrients to the low-lying wetland (Young et al. 2010). The upland part of the catchment including the plateau, stream valleys and slopes comprise the largest part of the catchment (78%). The lowland, consisting of two large lakes, wet meadows and tundra ponds of various dimensions, constitute 22% of the basin. The nearest government weather station is located at Resolute Bay, which lies approximately 145 km to the southwest (Figure 1).
Figure 1  (a) Location of PBP (75°40'N, 98°30'W), Bathurst Island, Nunavut, Canada (map modified after Young et al. 2010). Inset map shows the location of Bathurst Island in relation to other Canadian Arctic islands. (b) Detailed map of the PBP watershed including the lowland boundary, based on a detailed DEM (30 m) courtesy of Sina Muster, Alfred Wegner Institute and NTS maps (1:50,000) of the area.
Young & Labine (2010) indicate that PBP’s spring and summer climate is similar to Resolute Bay, hence it can be considered to have a polar desert climate (cool and moist). As yet, no evaluation has been made of PBP’s fall and winter climatology but expectations are that it is similar to that of Resolute Bay. Continuous permafrost underlies this extensive wetland area, and by the end of the summer the seasonal active layer reaches about 1.0 m in dry, gravelly areas and 0.3–0.5 m in wet, boggy ground (Young et al. 2010).

METHODS

Snowcover

Detailed end-of winter snowcover measurements (depth, density) were conducted from mid- to late May (2008–2010). A terrain-based snow survey of representative sites (plateau, wet meadow, pond, lake, late-lying snowbed, stream valley, etc.) followed after Woo (1997). This normally consisted of a series of transects laid out along representative terrain (plateau, late-lying snowbed, lake, wet meadow), with the beginning and end of transects marked by a Garmin GPS (±5 m). Depending on the length of each transect, snow depth measurements were made every 2, 5, 10 or 50 m. Each depth measurement was the average of four measurements in a square around the intended point. In all cases, snow depth (cm) was measured with a metal ruler (±0.5 cm) or, when the snowpack was deep, a steel rod and metric measuring tape (±0.5 cm). For ponds, four transects were located around the general pond border and two across the centre to encompass the pond and its adjacent catchment. Here, snow depth measurements were taken along each transect at the appropriate interval (1–5 m). For all terrain features, the snow density measurements (kg m⁻³) were taken at the beginning, middle and end of each transect using a Meteorological Service of Canada snow core (±7%). Additional measurements were made along the longer transects. Here they were usually taken every 10th snow depth measurement. Snow water equivalence (mm) for each terrain unit was calculated as:

$$\text{SWE}(i) = \frac{d_i \rho_i}{\rho_w}$$

where $d_i$ is the average snow depth of terrain type $i$ (mm), $\rho_i$ is average snow density (kg m⁻³) and $\rho_w$ is the density of water (1,000 kg m⁻³).

The PBP catchment basin, upland and lowland areas along with terrain subcategories including slope, aspect and average elevation were classified based on National Topographic System (NTS) maps of the area at the 1:50,000 scale and digital elevation model (DEM) images. At this particular map scale, only larger rivers and streams are visible and only these were included in the calculation of total valley channel area. Lengths of streams were determined in GIS using digitized NTS map products; widths were estimated based on visual interpretation of the NTS maps in combination with DEM imagery to calculate valley channel area. Locations and areas of late-lying snowbeds were identified based on contours lines. The remaining upland area pertaining to plateau terrain was calculated by subtracting the total area of the valley channels and the late-lying snowbeds from the total upland basin area. The lowland area was defined and calculated as an area below 60 m a.s.l.

Snowmelt

Snowmelt estimates at PBP consisted of both direct measurements of surface ablation of key terrain types using the approach outlined by Heron & Woo (1978) and the employment of a physically based snowmelt model (Woo & Young 2004). Specifically, ablation measurements were made daily at four sites: pond, wet meadow, late-lying snowbed and plateau. This involved measuring the distance (±5 mm) from the top of the snowpack to a stable reference point (i.e. a line held taut between two dowels). An average of 10 height measurements at each site was recorded daily, with measurements being made in the central part of the line away from pole edges. Daily surface density measurements ($n = 5$, each scoop was 200 cm³ in volume) were also recorded. Estimates of snow lowering together with density changes allowed surface melt in snow water equivalent units (mm day⁻¹) to be determined. In 2010, a series of three snow ablation poles were inserted into a deep valley snowpack site located c. 2 km from base camp. The distance (±0.5 cm) from the top of the pole to the snowcover was monitored every few days but not surface
density. These snowpack height changes, together with daily snow density values from the late-lying snowbed ablation site which likewise had a deep snowpack, provided an estimate of snowmelt.

The physically based snowcover model (Woo & Young 2004) was applied to distribute the snowcover and melt across the PBP watershed for 2008–2010. An overview of the modeling structure is found in the Appendix (available online at http://www.iwaponline.com/nh/044/083.pdf), while full details on the methodology and algorithms pertaining to the model are in Woo & Young (2004).

Here we provide a brief summary. Initially, this model builds up the snowcover at varying terrain units with respect to a base station. In this study, the base station was a centrally located wet meadow site. The model considers slope, aspect and lapse rates when determining initial snowcover depth and cold content for the different terrain units. Snowmelt is based on the surface energy balance framework, in which the energy available for melt \( Q_m \) can be broken down into the following components (Woo & Young 2004):

\[
Q_m = Q^* + Q_h + Q_e + Q_p
\]

(2)

where \( Q_m \) is the heat available for melt, \( Q^* \) is net radiation, \( Q_h \) is the sensible heat flux, \( Q_e \) is the latent heat flux and \( Q_p \) is the energy from rain during snowmelt. Net radiation was calculated by partitioning it into its shortwave and long-wave components. The turbulent fluxes \( Q_h, Q_e \) were estimated using the bulk transfer method (Price & Dunne 1976), an approach found suitable for High Arctic sites (Woo & Guan 2006; Abnizova & Young 2010). The energy available for \( Q_m \) can be converted into melt \( (M) \) by:

\[
M = \begin{cases} 
\frac{Q_m}{\lambda \rho_w} & \text{if } Q_m > 0 \text{ and cold content is } 0 \\
0 & \text{otherwise} 
\end{cases}
\]

(3)

where \( \lambda \) is the latent heat of fusion and \( \rho_w \) is the density of water.

Meteorological inputs to the model include hourly inputs of incoming solar radiation \( (K \downarrow , W \ m^{-2}) \), air temperature \( (T_{air}, \ ^\circ C) \), relative humidity \( (%) \), precipitation \( (mm) \) and atmospheric pressure \( (P_0) \). Incoming radiation \( (K\downarrow) \) is adjusted for slope and aspect, and adjustments can be made to the albedo algorithm for different snowpacks (seasonal versus persistent, such as in the case of late-lying snowbeds). The albedo decay follows the empirical function established by Woo & Dubreuil (1985):

\[
\alpha(\tau) = a_0 - \frac{a_0 - a_b}{\exp(B_1 + B_2 \tau)}
\]

(4)

where \( a(\tau) \) is average daily albedo on day \( \tau \) after the initiation of melt, \( a_0 \) is the albedo on the first day of melt and \( a_b \) is the albedo of the bare ground. The values of \( a_0 = 0.8 \) and \( a_b = 0.2 \) often describe the initial and final albedos of the snow site. For a flat site, Woo & Young (2004) found that \( B_1 \) and \( B_2 \) were equivalent to 2.46 and \(-0.26 \). However, for deep valley snow where snow lingered longer, the albedo decay was more gradual and \( B_2 \) had a value of \(-0.07 \). Whenever new snow falls, the albedo rises back to \( a_0 \) but after this fresh snow is melted, the exposed old surface albedo will be that value before the snowfall and the day counter \( \tau \) will resume, continuing the albedo decline. Adjustments can also be made to the wind function. Woo & Young (2004) found that the ratio of windspeed between the base (flat site) to a plateau, valley and northwest slope at the McMaster Basin was 1.09, 0.96 and 0.85. At PBP, analysis of wind data (valley versus a flat open site) in 2010 did not show any significant variation in windspeed, hence no adjustment to the winds was made in the model.

Meteorological information for the model came from an automatic weather station (AWS) near the base station (see Young & Labine 2010 for details of the instrumentation used). A Hobo pressure transducer provided atmospheric pressure data and, in the case of missing data, it was retrieved from the government weather station at Resolute Bay. The model outputs daily melt in snow water equivalent terms \( (mm) \) for various terrain units. The reliability of the model for this site along with others has been well documented (Young 2008; Abnizova & Young 2010). Assini & Young (2012) applied it to assess the spatial pattern of melt across the low-lying wetland in 2008 and 2009, while Howell et al. (2012) confirmed its reliability against QuikSCAT satellite imagery. In this study, we utilize it to facilitate greater understanding of snowcover and melt energetics of upland sites in relation to the low-lying wetland landscape.
RESULTS AND DISCUSSION

Summary of climatic conditions

Climatic conditions at PBP during the period 2008–2010 (Figure 2) are comparable to those of Resolute Bay, confirming a polar desert climate designation (Table 2). Hourly air temperatures were alike in the early part of June for all years, but substantially increased during late June 2010 (Figure 2). Overall, during the main snowmelt season (June), air temperatures were slightly warmer in both 2008 and 2010 than in 2009 (Table 2).

Precipitation at PBP was negligible during snowmelt (Assini & Young 2012). No rain was recorded during the

![Figure 2](image-url) Seasonal climatology from the centrally located automatic weather station during snowmelt at PBP (2008–2010): (a) $K_1$ (W m$^{-2}$); (b) $T_{air}$ (°C); (c) $U$ (m s$^{-1}$); (d) RH (%); and (e) PPT (mm day$^{-1}$).
2010 study period except for drizzle on 10 June and a trace rain event (0.1 mm) on 12 June at the onset of snowmelt. Measured amounts of rain during the snowmelt period were similar in 2008 and 2009. Precipitation totals were comparable to Resolute Bay except in 2008 when this coastal station reported higher precipitation receipt in June (Table 2).

Mean windspeed was similar for the 3 years (4.0–4.3 m s⁻¹) and akin to Resolute Bay (Table 2). Wind speeds were moderate in 2010 but, in early June 2008 and 2009, they occasionally reached 11.0 and 9.4 m s⁻¹, respectively (Figure 2). Likewise, there was little difference in relative humidity (%) for all 3 years, which demonstrated values over the range 77–81% during June. Due to long stretches of low cloud, incoming shortwave radiation receipt in 2008 and 2009 (K↓) at PBP was similar for June (2008: 714 MJ m⁻² versus 2009: 706 MJ m⁻²) and the often overcast days triggered cool air temperatures. This pattern differed from 2010 when an extended stretch of clear and sunny weather enhanced K↓ inputs (852 MJ m⁻²; Figure 2).

**End-of-winter snowcover**

Figure 3 illustrates the characteristics of upland and lowland winter snowpacks. Aside from the windswept plateau areas with their shallow snowpacks and a hard surface layer, other typical upslope terrain such as incised stream valleys and the lee of slopes, sites of late-lying snowbeds accumulate much snow. Stratigraphy analysis of deep snowpits in upland areas (2009, 2010) reveal multiple layers of snow with varying grain size and density than in shallow upland (plateau) and low-lying (wet meadow, pond) snowpits. Deeper snowpacks of upland areas also had colder temperatures than lowland sites.

Figure 4 illustrates the end-of-winter snowcover survey results for the various years (2008–2010) and includes snow depth (cm), snow density (kg m⁻³) and snow water equivalent (mm) for representative terrain units comprising the upland areas (plateau, valleys) and the low-lying wetland (ponds, wet meadow, lake). As revealed by Figure 3 (snowpit stratigraphy), the windblown plateau and lowland sites (ponds, wet meadow, lake) captured less snow than upland sheltered stream valleys and the lee of slopes. Variability in snowcover remains the norm across the years with the deepest snowpack occurring in 2010 and the shallowest in 2009. As expected, differences in snow water equivalence result from variations in snow depth (6–91 cm) rather than density. Snow density was more uniform across most terrain types, ranging 200–300 kg m⁻³. Higher densities occurred in deep snow (valleys, lee of slopes) and/or near the surface of shallow snowpacks where strong winds enhanced wind slab processes (e.g. plateau; Figure 4(a)).

Ponds located in the north and eastern sectors of the pass had lower densities than the western and southern situated ponds. This suggests that they were protected by nearby hills which were effective in blocking strong winds, which blew predominantly from the north (Figure 4(b); Young & Labine 2010).

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**Table 2** Summary of climatic conditions at Polar Bear Pass (Central wetland weather station) and the government weather station at Resolute Bay, Nunavut, June 2008–2010. At Polar Bear Pass winds were measured at c. 2.0 m (central wetland site) and at the government weather station (Resolute Bay) they are measured at 10 m. Values in brackets indicate estimated windspeeds at 2.0 m at Resolute Bay. Here a surface roughness value of 0.001 m typical of a smooth snowcover is assumed (Oke 1987). N/A indicates that data are not available.

<table>
<thead>
<tr>
<th>Location</th>
<th>Mean Tair (°C)</th>
<th>Mean (RH%)</th>
<th>Mean U (m s⁻¹)</th>
<th>Total PPT (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Polar Bear Pass</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2008</td>
<td>2.6</td>
<td>81</td>
<td>4.3</td>
<td>8.7</td>
</tr>
<tr>
<td>2009</td>
<td>1.9</td>
<td>77</td>
<td>4.0</td>
<td>2.1</td>
</tr>
<tr>
<td>2010</td>
<td>2.6</td>
<td>80</td>
<td>4.0</td>
<td>0.0</td>
</tr>
<tr>
<td><strong>Resolute Bay</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2008</td>
<td>2.2</td>
<td>N/A</td>
<td>7.0 (5.7)</td>
<td>21.6</td>
</tr>
<tr>
<td>2009</td>
<td>1.0</td>
<td>N/A</td>
<td>5.7 (4.8)</td>
<td>3.4</td>
</tr>
<tr>
<td>2010</td>
<td>2.2</td>
<td>N/A</td>
<td>4.9 (4.0)</td>
<td>3.4</td>
</tr>
</tbody>
</table>
Figure 3 | Analysis of snowpits dug at the end of winter at PBP. Upslope: (a) plateau, 2009; (b) late-lying snowbed, 2009; and (c) deep valley, 2010. Lowland: (d) pond areas, 2009; and (e) wet meadow, 2009.

(continued)
Snowmelt

In High Arctic regions, the snow that has accumulated over the winter is rapidly melted in a short period in the spring (Woo & Young 2006). Snowmelt is a significant component of the annual water budget and this period generally produces the highest seasonal discharges, often leading to the highest water tables in wetland ponds and lakes. It is also the primary season when nutrients are transported into ponds, lakes and streams from upslope areas (both wetland and non-wetland terrain; Abnizova et al. 2012). The duration and end of the spring snowmelt period initiates other important hydrological processes such as ground thaw and evaporation. In addition, it normally entails a shift from surface to subsurface flow from upland areas into low-lying landscapes (Woo & Guan 2006).

Surface energy fluxes (upland versus lowland)

Figure 5 provides an indication of the modeled daily average energy fluxes to the snowpack for both upland and lowland sites during the main melt period (5–20 June); here, energy fluxes from 2010 are used as an example. In order to assess melt energy between years, seasonal totals of surface energy fluxes for the same time frame (5–20 June) for all years are listed in Table 3.

Due to limited rainfall in 2010, all of the melt energy was partitioned between \( Q^p \), \( Q_h \), and \( Q_e \) (Figure 5). Latent heat losses of 50 W m\(^{-2}\) per day occurred in early June at most sites and then shifted to small gains during steady and sizeable melt. The lowland wet meadow site had the earliest and sizeable gains of net radiation reaching about 250 W m\(^{-2}\). Around June 15, both this site and the upland plateau shared similar daily energy fluxes. Field
observations suggest that these two sites melted out by June 17 while the model suggests that it occurred 2 days later. The discrepancy in the melt-out dates (measured versus modeled) is reasonable given that direct snow ablation measurements are usually stopped when ablation poles start to lean, despite thin snow still remaining on the ground. For deeper snowpacks (i.e. late-lying snowbed sites), ablation poles are often re-inserted after they melt out in one location so that measurements can be extended.

In comparison to the plateau and the low-lying wet meadow, daily average energy fluxes, including net radiation at the late-lying snowbed, were dampened and did not rise above 150 W m$^{-2}$. This pattern reflects the initial site conditions of the model. Here, a thicker snowpack would

![Figure 4](https://iwaponline.com/hr/article-pdf/44/1/2/370179/2.pdf)
Figure 5 | Daily average energy fluxes during the 2010 melt season for representative upland sites: (a) plateau, (b) late-lying snowbed versus lowland terrain: (c) wet meadow. Positive fluxes of sensible and latent heat are gains to the snowbed, while negative values are losses. For graphing purposes, melt fluxes have been reversed.
Table 3  Modeled upstore/lowland surface energy flux totals (MJm⁻²) during the main melt period, 5–20 June, for 2008–2010. Numbers in parentheses indicate observed and modeled dates in June when initial measured average snowpacks (SWE, mm) melted out

| Modeled upslope/lowland surface energy flux totals (MJm⁻²) during the main melt period, 5–20 June, for 2008–2010. Numbers in parentheses indicate observed and modeled dates in June when initial measured average snowpacks (SWE, mm) melted out |
|---|---|---|
| **Plateau** | 2008 | 2009 | 2010 |
| Q^a | 3 | 3 | 156 |
| Q_N | 50 | 50 | 35 |
| Q_S | −28 | −36 | −20 |
| Q_M | 3 | 0 | 0 |
| Q_in | 55 (15/19) | 48 (11/22) | 179 (17/19) |
| **Late-lying snowbed** |  |  |  |
| Q^a | 27 | 27 | 31 |
| Q_N | 53 | 51 | 37 |
| Q_S | −27 | −35 | −19 |
| Q_M | 3 | 0 | 0 |
| Q_in | 58 (1/23) | 45 (13/28) | 51 (20/23) |
| **Wet meadow** |  |  |  |
| Q^a | 58 | 37 | 220 |
| Q_N | 57 | 56 | 40 |
| Q_S | −25 | −32 | −18 |
| Q_M | 3 | 0 | 0 |
| Q_in | 94 (13/18) | 64 (12/18) | 243 (17/19) |

*a* Indicates that ablation measurements ended 17 June. Late-lying snowpack had lost an estimated 90 mm of melt; the initial average snowpack was determined to hold 130 mm.

*b* Direct ablation measurements in 2009 were initiated on 26 May. By 1 June, the plateau had gained 7 mm SWE, the late-lying snowbed had lost 7 mm and the wet meadow had lost 4 mm. Dust on the late-lying snowbed was observed during 27–29 June and this likely accelerated observed melt (Adhikary et al. 2002).

have had a larger cold content to overcome prior to initiation of melt. In addition, the albedo algorithm appropriate for persistent snowbeds was selected for this site. This is the albedo signal which declines less rapidly than a typical snowpack (Woo & Young 2004), leading ultimately to lower receipts of net radiation in comparison to shallower sites. Field observations at the late-lying snowbed showed that its average snowpack depth melted out by 20 June, while the model suggested it occurred 3 days later (Figure 6).

According to Table 3, daily average fluxes were much lower for 2008 and 2009 than in 2010, due to cooler and more frequent cloudy conditions (see Figure 2). Net radiation never exceeded 150 W m⁻² in 2008 or in 2009. Comparable to 2010, the wet meadow site in 2008 exhibited the largest positive fluxes followed by the plateau and then the late-lying snowbed. In 2009 however, the energy fluxes were quite similar across upslope and lowland sites. This might be partially explained by the cooler and cloudy conditions in 2009.

It is interesting that the discrepancy between measured and modeled melt was greater during cool and cloudy years (2008, 2009) than in a warm and clear melt season (2010; Table 3; Figure 6). For example, in 2009 a significant difference in melt (measured versus modeled) at the late-lying snowbed can be attributed to strong winds which carried dust from the adjacent plateau to the snowbed (Field note observations, June 2009). This thin layer of sediment most likely accelerated snowmelt, and is a phenomenon which has been frequently reported by others (Drake 1981; Woo & Dubreuil 1985; Young & Lewkowicz 1990; Ledley & Pfirman 1997; Hadley et al. 2010). Unfortunately, this process of dust deposition and its impact on albedo is not considered in Woo & Young’s (2004) snowmelt model or in other hydrologic models developed for cold regions (e.g. CRHM: Pomeroy et al. 2007; GEOTop: Endrizzi & Marsh 2010).

Like others, Adhikary et al. (2002) show that a thin dust layer over a snowpack can accelerate melt by decreasing the surface albedo and enhancing absorption of incoming radiation. Over short time intervals it can elevate surface temperatures (>0°C), thereby modifying sensible and latent heat fluxes to/from the snowpack. Conversely, thicker sediment (>a few centimetres) can insulate the underlying snowpack and thereby delay melt. Adhikary et al. (2002) acknowledge that surface albedo is a crucial parameter to determine snow ablation, and only fairly accurate albedo data on a spatial basis would yield good results. It is not practical, however, to acquire albedo data from relatively large, remote and inhospitable areas by ground-based measurements. To overcome this problem, high-resolution satellite imagery is the best option to support estimation of surface albedo. However, widespread utilization of it is still limited in High Arctic wetland studies due to coarse resolution (e.g. 1 × 1 km² of Moderate Resolution Imaging Spectroradiometer or MODIS), slow return time and interference with low-lying clouds, especially for optical types (e.g. Advanced Very High Resolution Radiometer or AVHRR; LandsAT; Hall et al. 2006).

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The daily performance of the model in 2010 in terms of evolution of snow water equivalence (mm) is provided in Figure 6. Similar diagrams (measured versus modeled) have recently been reported for both 2008 and 2009 (Assini & Young 2012; Howell et al. 2012), and have been alluded to here in Table 2. The model underestimates the start and end of melt for both upland and lowland sites, but in all cases the aerial surveys confirm modeled results since, as mentioned earlier, direct ablation measurements stop before the snow has completely disappeared from a site due to melt-out (leaning) of ablation poles. Likewise, as indicated earlier in this study, ablation poles were often moved and inserted into deeper snowpacks in order to extend surface ablation measurements. In relation to 2010, measurements at the

**Figure 6** SWE evolution (2010) of representative upland sites: (a) plateau, (b) late-lying snowbed and (c) valley snowpack; and lowland terrain: (d) pond 7 and (e) wet meadow. For both modeled and measured estimates, initial snow water equivalent of the snowpack is the average for that terrain unit. Direct snow ablation measurements at the late-lying snowbed continued until 30 June, while those at the valley snowpack extended until 1 July.

(continued)
late-lying snowbed continued until 30 June, while those at the valley snowpack were extended until 1 July.

PBP basin

Since no snow survey was conducted in the southern sector of the basin in 2008 (i.e. slopes, plateau, stream valley), no further analysis was conducted at the basin scale and only 2009 and 2010 basin results are reported here. To assess the temporal pattern of melt over the entire PBP watershed, including the upland and low-lying wetland, the end-of-winter snowpack information and daily rate of snowmelt pattern for both 2009 and 2010 was up-scaled to the entire basin using modeled melt and ArcGIS (Table 4; Figures 1 and 7). Overall, the lowland comprises 22% of the basin and the upland 78%. Within the lowland, the ponds and lakes share similar areas while the wet meadow zone comprises the largest terrain unit. The plateau area makes up the largest terrain type of the upland region (65%), while the steep slopes which shelter late-lying snowbeds comprise <10% here. Not surprisingly, the proportion of stream valleys in the uplands is small (c. 3%), though they tend to capture the most snow. For this modeling exercise, no differentiation was made between the north and southern plateaus and end-of-winter snowpacks and melt progression at these two locations were averaged. Since most of the stream valleys run north–south or south–north, only these drainage channels were categorized here. Similarly, in the lowland, end-of-winter snowpacks and melt progression were averaged across all ponds surveyed in the pass (n = 23). This differs from the study of Assini & Young (2012) which mapped spatial melt across the low-lying wetland and weighted pond accumulation and melt in the different sectors of the pass by area (north, south, east and west).

Figure 7 illustrates the snow water evolution of the entire lowland and upland regions in relation to the PBP watershed for 2009 and 2010. Figure 8 documents the snow-cover extent in the north-central part of the basin during aerial surveys; the photos generally support model results. The upland plateau and the low-lying ponds and wet
meadow of the northern part of the basin are generally free of snow while residual snow occurs in the lee of slopes and in stream valleys. Figure 8 indicates that snow persists in the southern part of the pass in both lowland and upland areas however, and reveals that this pattern was more evident in 2009. Assini & Young (2015) attribute this to a deeper and cleaner snowpack here and a cooler microclimate which causes a lag in melt compared to the northern sector. In comparison to other studies (Table 1), the lowland which comprised two large lakes, numerous ponds and an extensive wet meadow exhibited a shallow wetland snowpack in both 2009 and 2010. Snow storage here is indicative of polar oasis-type environments (warm/dry-limited precipitation; e.g. Glenn & Woo 1997; Yang & Woo 1999; Woo & Guan 2006), but not of wetlands experiencing a polar-desert climatic regime (cool/wet-heavy precipitation) which generally possess deeper snowpacks. This discrepancy between the snowpacks measured at PBP for all 3 years compared with other High Arctic wetland sites is not yet well understood.

In a shallow snowpack year (i.e. 2009) the lowland initially held more snow than the upland. This pattern differed from a year with a deeper snowpack (2010). This is not surprising given field observations of a thin plateau snowpack in 2009 (Figure 3). In 2009 and 2010, melt at the basin scale was controlled primarily by its progression in the upland regions (e.g. late-lying snowbeds and plateau). Despite a deeper snowpack in 2010, the pattern of melt (i.e. initiation, duration and end-date) between the lowland, upland and basin regions was quite similar. This relates to the extended period of sunny, warm weather in 2010 which accelerated melt in all areas both the uplands and in the lowlands. It is expected that under overcast and cooler conditions, the melt progression at the basin scale would have been quite similar to 2009 with both the duration and end-date of melt being extended. Similar findings are reported by Eaton & Wendler (1982) for a permafrost watershed in interior Alaska.

Snowcover results between the upland and lowland do suggest that the uplands might be an important source of water to the low-lying wetland given its much larger area and potential to store more snow. However, to better evaluate this contribution and to fully assess its importance to this wetland’s sustainability, we need to first analyse the spatial pattern of melt in the upland in relation to the lowland and the entire catchment. In addition, stream discharge from uplands to the lowlands has been estimated for 2008–2010 (Young et al. 2010), but a full assessment of hillslope–wetland runoff in relation to drainage from the entire PBP watershed will not be made until completion of the 2012 field season. Only then can we more accurately assess the role played by uplands in funneling water into the wetland and determine its sustainability in response to both present and future climate variability/change.

In summary, this study (which couples detailed fieldwork with a physically based modeling approach) contributes to an improved understanding of snowcover

<p>| Table 4 | Terrain areas (km² and %) of Polar Bear Pass watershed. Total basin area is c. 422.1 km². The upland area comprises 77.7% of the total watershed while the lowland approaches 22.3% |</p>
<table>
<thead>
<tr>
<th>Upland area (c. 328.1 km²)</th>
<th>Terrain area (km²)</th>
<th>% of upland area to total basin area</th>
</tr>
</thead>
<tbody>
<tr>
<td>N-facing slopes (snowbeds)</td>
<td>26.7</td>
<td>6.3</td>
</tr>
<tr>
<td>S-facing slopes (snowbeds)</td>
<td>13.4</td>
<td>3.2</td>
</tr>
<tr>
<td>N-facing stream valleys</td>
<td>4.9</td>
<td>1.2</td>
</tr>
<tr>
<td>S-facing stream valleys</td>
<td>7.4</td>
<td>1.8</td>
</tr>
<tr>
<td>Plateau</td>
<td>275.6</td>
<td>65.3</td>
</tr>
<tr>
<td>Lowland area (c. 94.0 km²)</td>
<td>Terrain area (km²)</td>
<td>% of lowland area to total basin area</td>
</tr>
<tr>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td>Lakes</td>
<td>7.4</td>
<td>1.8</td>
</tr>
<tr>
<td>Ponds</td>
<td>7.8</td>
<td>1.8</td>
</tr>
<tr>
<td>Wet meadow</td>
<td>78.8</td>
<td>18.7</td>
</tr>
</tbody>
</table>
and melt processes of an extensive low-gradient wetland located in the Canadian High Arctic. It is anticipated that it will advance global modeling efforts of future northern wetland decline (Avis et al. 2013), especially for models which do not evaluate inputs of water into wetlands from adjacent sloping terrain. These sources of lateral water inputs into low-lying wetlands both during and after seasonal snowmelt might be critical in offsetting future impacts of thawing permafrost.

**CONCLUSIONS**

Snow is an important source of water to arctic ecosystems, allowing ponds, lakes and wet meadows to be refreshed and recharged. Our results suggest that snowcover across the low-lying wetland, which comprises 22% of the PBP watershed (422 km²), is generally low (ranging 44–66 mm) and more typical of wetlands experiencing a polar oasis-type climate (Woo & Guan 2006)
than a polar desert climate (Abnizova & Young 2010). Modeled and measured snowmelt estimates are reasonable for both lowland (wet meadow, pond) and upland sites (plateau, late-lying snowbed, deep valley stream), although the model does tend to underestimate the initiation of melt and its duration. End-dates of melt are similar except for those in 2009. The discrepancies in the results reflect errors in measurement (e.g. early melt-out of ablation poles before the snowpack is completely gone) and model errors. In the latter case in 2009, the albedo algorithm in the model did not take into account enhanced aeolian deposition on the late-lying snowbank which lowered the reflectivity of the snowpack and accelerated snowmelt (Adhikary et al. 2002).

Overall, snowmelt in all 3 years began in June and was slow for a few days until air temperatures rose above 0°C, at which time active melt occurred and the snowcover in the lowlands was depleted by the third week in June. Net radiation fluxes dominated in warm sunny seasons while sensible heat fluxes to the snowpack accelerated snowmelt in cooler, overcast conditions. Deeper snow can linger in the lee of slopes or incised stream valleys of the upland region, providing additional meltwater and nutrients to nearby ponds and wet meadows after the seasonal snowpack has been depleted (Young et al. 2010).

Future work will explore the spatial pattern of snowcover and melt across this large wetland system (both upland/lowland) and other large extensive wetlands (i.e. southwest Bathurst Island). This information is required before we can fully assess the timing and magnitude of snowmelt discharge into coastal polar waters, and fully determine the importance of uplands in transferring water to adjacent low-lying wetlands (both surface/subsurface flow).

**ACKNOWLEDGEMENTS**

This research is supported by the Natural Sciences and Engineering Research Council (NSERC) of Canada, the Federal Government of Canada International Polar Year program and the Network Centre of Excellence (ArcticNet). We are grateful for the excellent logistical support from Polar Continental Shelf Program (PCSP) and Northern Student Training Program (NSTP). Many thanks to A. Croft, A. H. McGregor, N. DeMiranda, J. Siferd and A. Swidzinski for their assistance in the field. We wish to thank Sina Muster, PhD student, Alfred Wegner Institute, Potsdam, Germany for providing the detailed DEM of PBP. It was used in this study to construct the site map which was carefully produced by Carolyn King. Professors Ming-ko Woo and Lars Bengtsson provided useful comments on an earlier draft of this paper, for which we are grateful.
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First received 9 November 2011; accepted in revised form 29 May 2012. Available online 31 July 2012.