Swings in runoff at Polar Bear Pass: an extensive low-gradient wetland, Bathurst Island, Canada

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ABSTRACT

Hydrologic studies in northern landscapes indicate there is a critical need to explore how arctic stream discharge patterns and water budgets may be shifting in response to climate warming. The focus of this study was to: (1) assess the pattern of runoff out of Polar Bear Pass, a low-gradient watershed (75°40′N, 98°30′W), during two contrasting spring/summer seasons: 2012 (warm, early melt) versus 2013 (cool, late melt); (2) quantify the seasonal water budgets; and (3) place these results in the context of other arctic basin studies. The end-of-winter snowpack was quantified using a terrain-based approach. A physically based snowmelt model using local weather station data provided daily melt estimates. Streamflow at the eastern outlet was estimated using the mid-section velocity approach. Snow water equivalent (SWE) was higher in 2013 while snowmelt began and ended earlier in 2012. Stream hydrographs showed a rapid rise in flow driven by meltwater from the northern part of the Pass in 2012. This was followed by a series of secondary peaks, melt contributions from the southern end. In 2013, the largest runoff peaks came from the southern sector. Runoff ratios and water budgets varied between the two years, and runoff in 2013 was similar to High Arctic watersheds in the early 1970s.

Key words | arctic snow cover, arctic wetland, runoff ratio, snowmelt, streamflow

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INTRODUCTION

Wetlands are critical landscapes in the High Arctic, providing food for migratory birds and larger fauna such as caribou and muskoxen. They also serve to store and replenish freshwater supplies and recently they have attracted interest in terms of their role in greenhouse gas exchange (Meng et al. 2016). Snow remains an important source of water into these ecosystems, often replenishing ponds and lakes and re-saturating wet meadow areas at the end of a cold winter season. Evaporation is a key loss of water from these wetlands during the spring and summer seasons (e.g., Bowling & Lettenmaier 2010; Lilijedahl et al. 2011) but runoff out of these wetlands into coastal arctic waters is less well known (Fortier & Allard 2007). Hence, our ability to evaluate the flow of carbon, sediments and nutrients out of these wetlands is also limited.

Recently, streamflow studies in Northern Canada (Déry & Wood 2005; Déry et al. 2009) have shown an 'intensification' of runoff patterns such as earlier freshets, flashier peak flows and increases in baseflow in response to climate warming, but understanding of these changes in Canada's most northern basins, especially ones dominated by wetlands, is limited. Laudon et al. (2017) indicates that such changes can often be abrupt and unexpected, as catchment storage and release of water is inherently nonlinear. A better understanding of these ecosystems is essential in order to sustain them, especially in a rapidly changing region. Karlsson et al. (2015) similarly note the growing interest in arctic river discharge as shifts in the runoff pattern can signal alterations in the terrestrial hydrological cycle.

It is not yet apparent whether recent climate warming in the Eastern Canadian High Arctic (Woo & Young 2014) is translating into flow regime shifts for streams draining extensive wetland systems located here. Therefore, the objectives of this study are to: (1) assess the pattern of runoff out of the eastern sector of Polar Bear Pass (PBP), an extensive low-gradient wetland, during two contrasting spring/summer seasons: 2012 (warm, early melt) versus 2013 (cool, late melt); (2) quantify the seasonal water budgets for these two contrasting seasons; and (3) place these results in the context of other arctic watershed studies both in the Canadian High Arctic and elsewhere.

STUDY AREA

The study took place at PBP, which is located in the middle of Bathurst Island (75°40′N, 98°30′W), during two spring/summer seasons: 2012 and 2013 (Figure 1). The nearest government weather station to PBP is Resolute – Qausuittuq, Cornwallis Island (67.7 m a.s.l. 74°43′N, 94°58′W) (Figure 1). This weather station (ongoing since 1947) is considered to have a polar desert climate (Woo & Young 1997). In previous years at PBP, a range of hydrology, biogeochemistry, climatological and remote sensing studies have been conducted (e.g., Young & Labine 2010; Young *et al.* 2010,
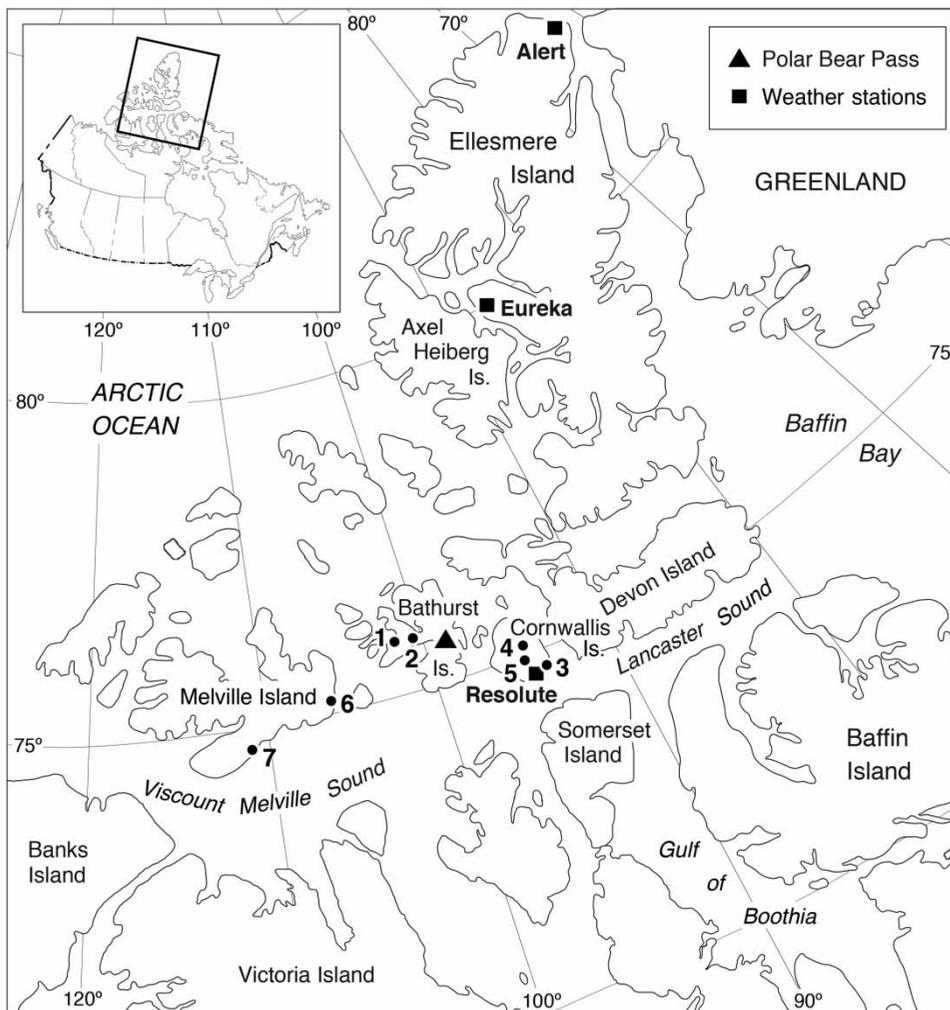

Figure 1 | Location of Polar Bear Pass, Bathurst Island, Nunavut, Canada (75° 40′N, 98° 30′W). Small watersheds mentioned in the paper are also indicated: 1, Snowbird Ck.; 2, Whitebear Ck.; 3, Mecham River; 4, Allen River; 5, McMaster River; 6, Consett Head River; 7, Cape Bounty. Additional details about these basins can be found in Young *et al.* (2015).

2017; Abnizova *et al.* 2014; Muster *et al.* 2015) but no stream-flow studies have been conducted at the basin outlets. PBP has a watershed area of 422.1 km², and is a designated National Wildlife Sanctuary and Ramsar site – a wetland of international importance as designated by the United Nations. The low-lying wetland (22.3% of the total basin area) runs east–west and, as is typical of wetland complexes, comprises two large lakes, a mosaic of ponds, and areas of wet and dry ground. Wet meadows here are lush, having a rich cover of grasses and sedges, often with *Salix arctica* hummocks. Hillslopes and barren uplands (77.7% of the area) border the wetland rising from about 23 m up to about 150 m a.s.l. V-shaped stream valleys (about 50 of varying order) notch the bordering hillslopes and are effective in transferring water and nutrients into the low-lying wetland (Young *et al.* 2010; Abnizova *et al.* 2014). Late-lying snow-beds typically occur in the lee of slopes and in the incised valleys after the main snowmelt period has finished. Two large rivers cut through the wetland at its periphery. The Goodsir River, a braided gravel stream, flows from the north, transitioning into a well-defined channel as it moves east into Goodsir Inlet. Field observations indicate that it does not flood the wetland in the spring but runoff from the far eastern sector of the wetland drains into the Goodsir River via rivulets. The Caledonian River (not named, Figure 2) is a well-incised stream which flows north and then heads west where it joins the western outlet of the wet-land (as marked) to empty into Bracebridge Inlet. It also does not flood the wetland during spring melt but does drain the western sector of the wetland (Figure 2). Figure 2(b) provides an aerial view of the wetland while Figure 2(c) is a photograph of the eastern sector outlet. It is estimated to drain an area of 102.6 km², and is the focus of this particular study.

METHODS AND DATA

Snow cover

Detailed end-of-winter snow cover measurements (depth, density) were carried out from mid- to late May in each year. A terrain-based snow survey of representative sites (plateau, wet meadow, pond, lake, late-lying snowbed,

stream valley, etc.) was followed (see Woo 1997), with a likely mean error approaching 15% (Woo & Marsh 1978). The results of the local snow survey were then up-scaled to the regional scale utilizing two topographical maps covering the area (Caledonian River, 1985, 68H/11, 1:50,000 and McDougall Sound, 1994, 68H, 1:250,000). A 100 × 100 m grid was applied over the digitized maps, and each 1,000 m² grid cell (equivalent in area to a small- to medium-sized pond) was classified as one of the main terrain types. There were 25 possible values, based on terrain type and geographical location in the eastern sector of PBP, although four terrain types made up almost 80% of the area (Plateau-(N); Plateau-(S); Wet Meadow; and South Slope). This allowed snow survey estimates to be interpolated across this part of the watershed.

Snowmelt

Snowmelt measurements at PBP consisted of both direct measurements of surface ablation of key terrain types (e.g., pond, wet meadow, late-lying snowbed, plateau, etc.) using the approach outlined by Heron & Woo (1978) and employment of a physically based snowmelt model (see Woo & Young 2004). Specifically, direct measurements of ablation 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 string held taut between two dowels). Ten height measurements at each site were recorded daily and then averaged. Daily surface density was also recorded in order to determine surface melt in snow water equivalent units (mm/day) (see Young *et al.* 2010).

The physically based snow cover model (Woo & Young 2004) was useful in distributing the snow cover and computed daily melt (in mm) across the eastern sector of the watershed for two contrasting seasons: 2012 and 2013. Initially, this model builds up the snow cover at varying terrain units with respect to an index station – here, a centrally located wet meadow site. The model considers slope, aspect and lapse rates when determining initial snow cover depth and cold content for different terrain units (e.g., plateau, valley, wet meadow, pond, late-lying snowbed). Meteorological inputs into the snow model include hourly incoming
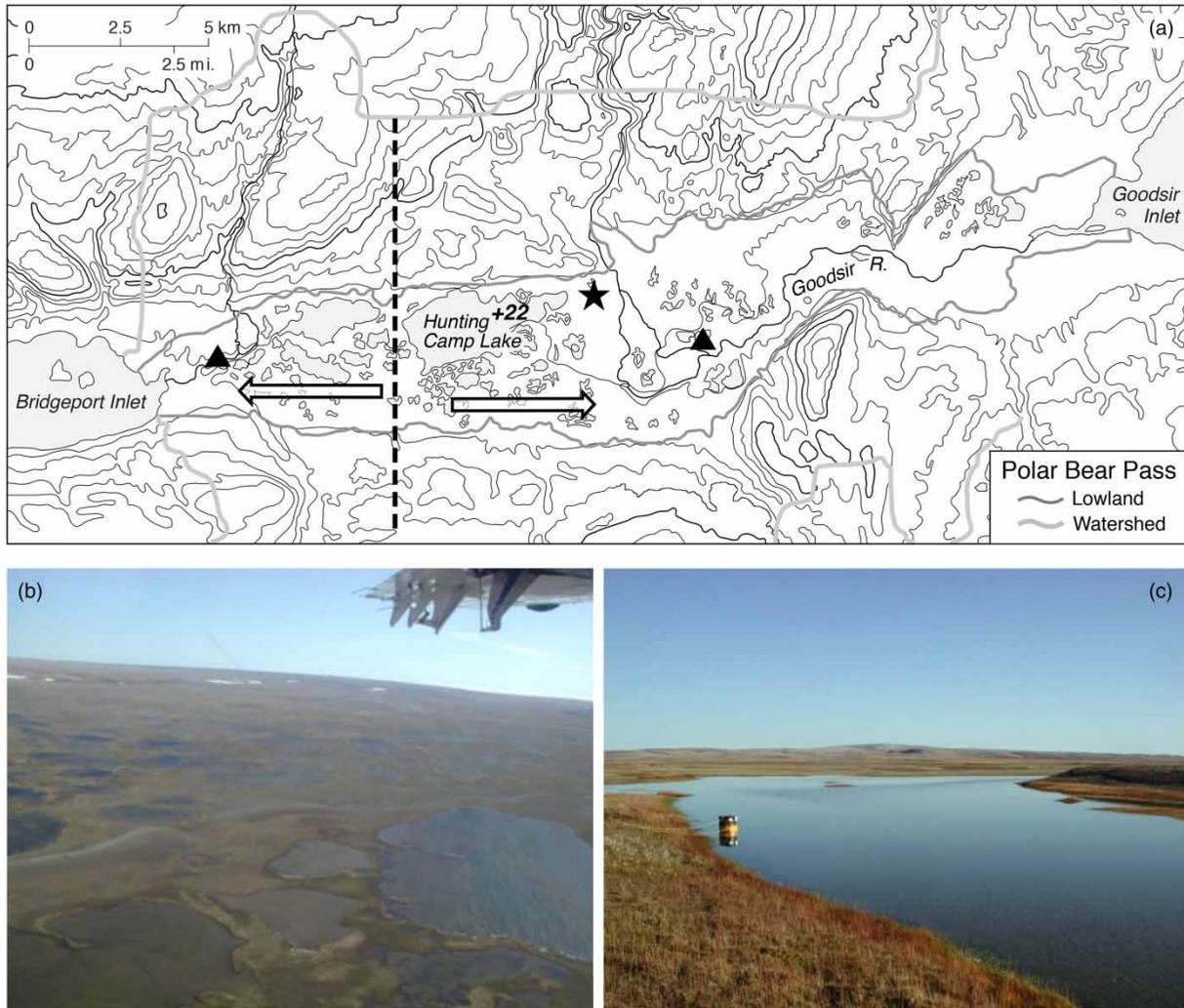
Figure 2 | Detailed map of PBP study watershed (a) including wetland and upland region. Elevation of Hunting Camp Lake is 22 m. a.s.l. and contours are 10 m (*Source:* Young *et al.* 2013). The vertical dashed line signals the 'likely' drainage divide in the watershed. The eastern sector drainage area is estimated to be 102.6 km². The star indicates the wetland Automatic Weather Station (AWS). Flow direction (arrows) and wetland outlets (triangles) are also indicated. Aerial view of the wetland (b) and photograph of the eastern outlet (gauging location) (c) taken on 4 July, 2012 are also provided.

solar radiation ($K\downarrow$, W/m²), air temperature (Tair, °C), relative humidity (%), wind speed (m/s), precipitation (R, rain or snow) (mm), and atmospheric pressure (Pa) (see Supplementary Figure 1 in Young *et al.* (2018)). Radiation is adjusted for aspect, temperatures and relative humidity are adjusted for elevation, and melt, here defined as ablation of the snowpack (mm, equivalent), will only occur once the daily cold content of the snowpack (mm, equivalent) is fulfilled. A Hobo pressure transducer provided hourly air pressure data (Pa). Adjustments can be made to the albedo algorithm in the model for different snowpacks (seasonal

vs. semi-permanent) as can the wind function (sheltered vs. exposed landscapes). Melt due to precipitation is adjusted for in the snowpack. If the air temperature is less than 0 °C, snow accumulates, and the cold content of the snowpack increases. Conversely, if the temperature is >0 °C, the Q_P melt flux (heat) due to rain-on-snowmelt is added to the snow cover. The turbulent fluxes, Q_H and Q_E, are calculated using the bulk transfer equations (see Price & Dunne 1976). Meteorological information for the model came from the centrally located wetland AWS (CAWS) and the main AWS (PAWS) situated on the Plateau

<1 km away (see Supplementary Table 1 in Young *et al.* (2018) for details on the instrumentation used). The model outputs daily melt (mm) for the various terrain units.

The snow model has shown its reliability in simulating daily snowmelt at PBP and other diverse terrain across the Canadian High Arctic, although it can over/underestimate melt for different terrain by a few days (e.g., Woo & Young 2004; Young *et al.* 2013). This is a limitation shared by other snowmelt models in arctic landscapes (e.g., Pedersen *et al.* 2016) and can be related to errors and assumptions inherent in the snow model and fieldwork results. Young *et al.* (2013) provide a thorough assessment of the limitations of the snowmelt model used in this study. Despite its shortcoming, it does employ easily accessible weather station data, and it is employed here to help shed light on how runoff out of the eastern sector of PBP is related to the spatial pattern of snowmelt between two distinct snow years.

Post-snowmelt (summer precipitation, evaporation)

In the post-snowmelt period (2012, 2013) precipitation, now defined as summer precipitation (R – rainfall and snow, mm), was measured at the CAWS using a Hobo tipping bucket rain-gauge (orifice at 0.19 m height; 0.2 mm per tip) and summed over the day. At the PAWS, the same device was used to measure summer precipitation in 2012 but it was replaced in 2013 by a Campbell Scientific Canada, Incorporated (CS 700) tipping bucket rain-gauge (orifice at 1.0 m; 0.2 mm per tip) (see Supplementary Table 2 in Young *et al.* (2018)). Summer precipitation was not corrected for wind errors, given that these estimates were made close to the ground surface where wind effects would be reduced. Nevertheless, we do recognize that this could be a source of error in our water budget estimates (about 7%), although errors for liquid precipitation are considered less than solid precipitation (e.g., Førland & Hanssen-Bauer 2000). Hourly evaporation (summed over the day) from the ponds, wetland areas and saturated ground following snowmelt, were estimated using the Priestley–Taylor (P-T) approach (Priestley & Taylor 1972), with $\alpha = 1.26$. Rosenberry *et al.* (2004) found that the P-T method compares well with other techniques in evaluating evaporation from wetlands. It is routinely used to estimate evaporation in High Arctic landscapes (e.g., Woo & Guan 2006; Abnizova *et al.* 2014). For moist and drying ground (e.g., polar desert terrain), the P-T approach was also employed, but instead, α was set as a function of near surface volumetric soil moisture (0–15 cm) (Marsh *et al.* 1981). Climate data employed to estimate hourly evaporation losses, such as net radiation, ground heat flux and air temperature, and relative humidity were obtained from CAWS and PAWS. Pond water heat flux was estimated using the approach outlined by Abnizova *et al.* (2014). Near-surface volumetric soil moisture (θ) was determined in both wet meadows and polar desert terrain using either continuously measuring Campbell Scientific TDR probes or Echo probes wired into Campbell Scientific dataloggers (see Young *et al.* 2015). In 2012, evaporation estimates were missing from polar desert terrain due to equipment failure (22 July–23 August). A regression equation of daily evaporation between wetland and polar desert terrain was used ($R^2 = 0.74$, $n = 46$) to fill in missing evaporation results. Finally, similar to snowmelt, summer precipitation and evaporation were areally weighted on a daily basis, in order to assess seasonal water budgets.

Table 1 | General climatic data at PBP plateau versus resolute, June–August 2012, 2013 and climate normals (1981–2010)

		1–30 June				1–31 July				1–31 August			
		Tair °C	PDD > 0 °C	R mm	U m/s	Tair °C	PDD >0 °C	R mm	U m/s	Tair °C	PDD >0 °C	R mm	U m/s
PBP	2012	3.7	114	0.2	4.2	8.7	269	27.0	4.1	3.4	106	14.8	3.5
	2013	−1.8	5	11.8	4.7	3.1	98	5.4	4.8	0.7	72	21.8	5.2
Resolute	2012	3.3	105	12.6	4.2	7.3	226	28.7	3.2	3.2	99	17.0	4.1
	2013	−1.2	12	34.2	4.5	3.2	96	21.4	4.0	0.9	59	16.4	4.2
	1981–2010	0.4	45	14.6	4.1	4.5	139	28.1	4.1	2.0	72	33.8	4.3

Tair is average air temperature; PDD refers to the number of positive degree-days >0 °C in the month; R refers to summer precipitation (rain, snow) totals for the month; and average wind speed (U) is adjusted to 1 m height, assuming a surface roughness (z_o) of 0.001 m.

Table 2 | Average snow survey and SWE index results for main terrain types at Polar Bear Pass

Terrain	2012				2013			
	Snow depth (mm)	Snow density (kg/m³)	SWE (mm)	SWE index (relative to wet meadow)	Snow depth (mm)	Snow density (kg/m³)	SWE (mm)	SWE index (relative to wet meadow)
Plateau (N)	63 (62)	252	16	0.50	140 (99)	314	44	0.76
Plateau (S)	331 (292)	313	99	3.09	286 (315)	315	90	1.55
Stream valley (N)	440 (534)	282	123	3.84	559 (496)	277	146	2.52
Stream valley (S)	815(598)	252	219	6.84	610(379)	316	194	3.34
Late-lying snowbed (N)	383 (245)	228	89	2.78	454 (419)	269	126	2.17
Late-lying snowbed (S)	399 (272)	242	101	3.16	312 (151)	280	86	1.48
Ponds (N)	136 (46)	218	31	0.97	217 (59)	246	55	0.95
Ponds (S)	229 (52)	273	62	1.94	304 (101)	252	77	1.33
Wet meadow	138 (34)	230	32	1.00	225 (50)	257	58	1.00
Polygon area	279 (50)	277	77	2.41	354 (228)	262	93	1.60
Hunting Camp Lake (HCL)	189 (97)	312	59	1.84	219 (74)	336	73	1.26

(N/S) denotes North/South part of the Eastern Sector. Values in brackets () are standard deviations of snow depth.

Streamflow

Seasonal runoff from the outlet of the eastern sector watershed into the Goodsir River (see Figure 2) was measured using the mid-width velocity approach. Stage (H) was monitored by recording water level sensors with Hobo pressure transducers (±3 mm) (Young *et al.* 2015). Frequent direct measurements of stilling well stages were made with a metric ruler (±5 mm). They provided an additional check on the reliability of continuous stage measurements at different time intervals, and/or were used to correct stages when the values drifted. Direct discharge measurements at both low and high flows were made to develop reliable stage-discharge rating curves for each year. Generally, due to shifting conditions in the channel, two equations were developed, one for an ice-filled channel and another for an ice-free channel. Field observations indicated that the basal ice in the stream did not melt *in situ*, but was instead released, lifted and carried into the main outlet channel (Goodsir River) within a short time – about a day. These equations with an R^2 typically >0.80 were then used to adjust the water stage into a continuous discharge record. In 2012, from 16 to 18 June, $Q = 24.60H^{3.2231}$, $R^2 = 0.99$, $n = 3$ (ice-

filled channel); from 19 to 24 June, until the end of the season $Q = 10.55H^{2.6461}$, $R^2 = 0.99$, $n = 6$ (ice-free channel). In 2013, from 26 June to 2 July, $Q = 17.096H^{3.3614}$, $R^2 = 0.99$, $n = 4$ (ice-filled channel); and from 2 July and onwards, $Q = 12.537H^{3.0878}$, $R^2 = 0.85$, $n = 10$ (ice-free channel). Errors in streamflow estimates can amount to 10 to 14% on average (Young *et al.* 2010).

RESULTS AND DISCUSSION

General weather conditions at PBP during the period from 2012 and 2013 are comparable to those of Resolute, confirming a polar desert climate designation (Table 1), although summer precipitation was much greater in 2013 at Resolute than PBP. Air temperatures and wind speeds at PBP were similar to Resolute in both years. Air temperatures fell above the long-term mean at Resolute in 2012 and below the long-term mean in 2013. Summer precipitation totals at PBP were comparable between the two years (2012–42 mm; 2013–39 mm), although the frequency distribution of the daily precipitation totals differed (data not shown). Resolute recorded more summer precipitation

than PBP in both years (2012–58 mm; 2013–72 mm) although the higher receipt in 2013 was below the long-term mean at Resolute (1982–2010). Positive degree days (June, July and August – JJA) for PBP were 490 in 2012 and 176 in 2013. For Resolute, estimates of JJA were 430 in 2012 and 169 in 2013 (see Table 1 for monthly totals). Determination of positive-degree days followed after Woo & Young (2014).

End-of-winter snow cover

Table 2 illustrates the end-of-winter snow cover results for 2012 and 2013 and includes snow depth (cm), snow density (kg/m^3) and snow water equivalent (mm) for typical terrain units comprising the PBP watershed. As expected, wind-blown terrain (elevated plateaus, frozen ponds and lakes) tend to accumulate less snow than sheltered valleys and the lee of slopes where winds are dampened (e.g., Woo et al. 1983; Yang & Woo 1999; Woo & Young 2004; Woo 2012: Table 4.1, p. 127). Like other environments, including temperate ones, snow cover variability in terms of water equivalent units largely depends on the variation in snow depth which ranges from 63 to >800 mm in 2012 and 140 to >600 mm in 2013. Snow density tends to be more uniform, ranging from 200 to 300 kg/m^3 for the terrain units in both years (Table 2). Higher densities occur in deep snow (valleys, lee of slopes) or shallow snowpacks (plateau) where strong winds enhance wind slab. Areally weighted snow (SWE, mm) across the eastern watershed sector was slightly higher in 2013 (81 mm) than in 2012 (72 mm) with the southern end generally capturing more snow (see Table 2). Surprisingly, the estimates for the wet meadow, ponds, lake and plateau areas (depths, density and SWE, mm) are akin to wetlands having a polar oasis-type climatic regime (Woo & Guan 2006) rather than ones influenced by a polar desert climate (Abnizova & Young 2010). Polar oasis-type regions (e.g., Truelove Lowland, Devon Island; Fosheim Peninsula, Ellesmere Island) are typically sheltered from poor weather (e.g., cloudy, cool and windy conditions) and often receive lower amounts of snowfall and rainfall than polar desert land-scapes. As well, they often experience warmer springs and summers due to elevated solar radiation levels (Edlund & Alt 1989; Edlund et al. 1990).

Snowmelt

Eastern watershed

Due to a deeper snowpack and a cooler spring in 2013, active snowmelt and ablation of the snowpack was strongly delayed in comparison to 2012 (Figure 3). In fact, snowpack ablation in 2013 did not begin until well after most of the snow cover had disappeared in 2012. As observed elsewhere in the Arctic, deeper snow lingers in the lee of slopes and in the stream valleys as the shallow snowpack across uplands, wet meadows, ponds and lakes disappears first. Aspect plays a defining role in the Pass, as the snowpack is slower to leave in the southern part of the Pass (north-facing). This snowmelt pattern is a regular occurrence each spring (Assini & Young 2012; Young et al. 2013) and in some years it is accelerated by aeolian processes. In 2009, strong north winds eroded the North plateau, blowing sediment onto the northern half of the Pass. This dirt accelerated melt here while the southern part of the Pass remained rela-tively clean. This erosion/deposition pattern failed to emerge in 2010, arising from a deeper and more extensive snowpack on the upland, implying that this phenomenon only occurs in low snow years when the North plateau is relatively barren or else blown free of snow by strong winter winds. The acceleration of melt due to aeolian ero-sion and deposition has been found elsewhere in the Canadian High Arctic (Lewkowicz & Young 1991; Woo et al. 1991).

Streamflow at PBP – 2012, 2013

Streamflow out of the eastern sector of PBP into Goodsir River was quite different in 2012 and 2013 owing to vari-ations in snow and climatic conditions (Figure 4). Initiation of streamflow was earlier and more dramatic in 2012, with the early peaks in runoff being driven by the rapid melt-out of the northern part of the Pass (Figure 3). A series of secondary peaks occurs when the southern part of the Pass (northern aspect) started to melt out (Figure 3). By the last week in June the outlet was already approaching baseflow conditions (lasting for 61 days) com-parable to several non-glacierized basins in Svalbard (Blaen et al. 2014). In 2013 streamflow did not begin until
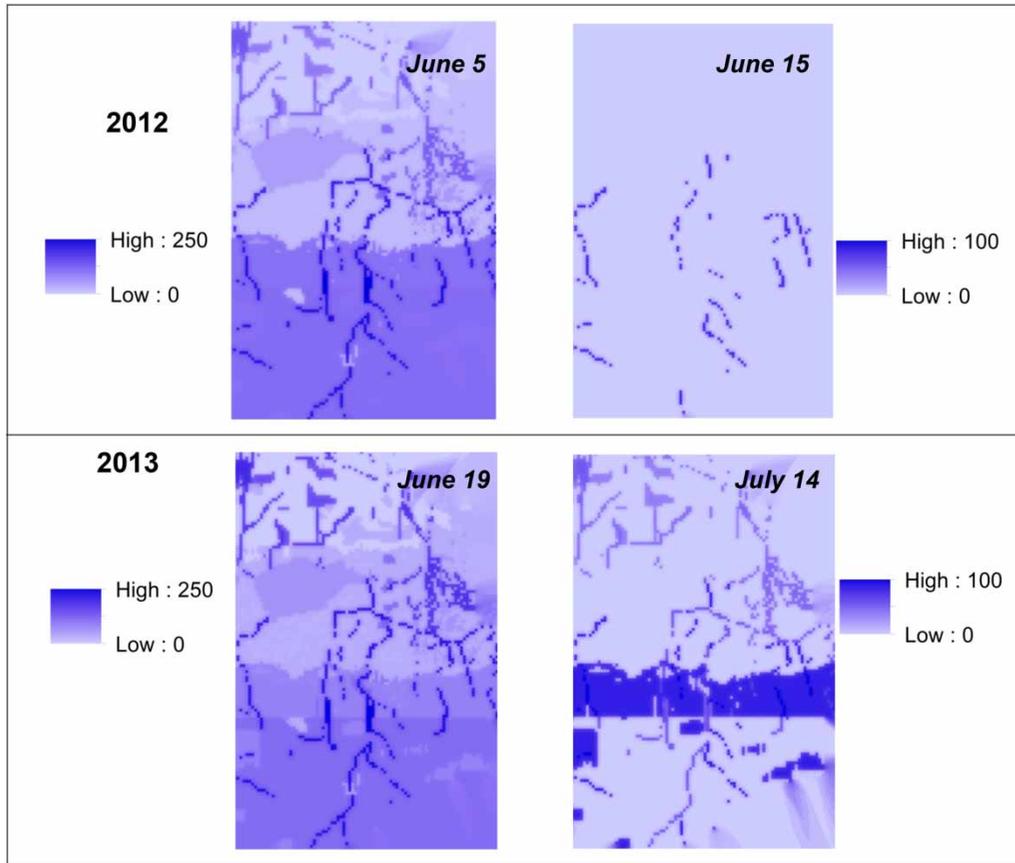
Figure 3 │ Modelled snowcover patterns (SWE, mm) for PBP (eastern sector) during the main snowmelt periods in 2012 and 2013. North is to the top of each image and the outline of Hunting Camp Lake (see Figure 2(a)) is visible in the left-hand plots.

late June, a delayed start considering that streamflow in 2012 was already at baseflow conditions by then. The 2013 streamflow hydrograph is different from 2012. The early start to the hydrograph is still being controlled by snowmelt coming from the north, but the biggest peaks in mid-July and the strong diurnal rhythm are indicative of active snowmelt (see Figure 3) from the southern part of the Pass. Baseflow also only lasted 17 days, although this could have been extended if not for logistical constraints in 2013. Despite the variations in timing and patterns of the hydrographs, an exceedance probability analysis of the runoff data (graphs not shown) revealed that streamflow out of the PBP eastern sector watershed is characterized by a nival-type regime (snowmelt dominated).

Runoff ratios differed between the two years. In 2012 the freshet Q/(Snowmelt + R) (5–23 June) was 66% and the seasonal Q/(Snowmelt + R) was 88% (5 June–23

August), while in 2013 the ratios were higher with the freshet Q/(Snowmelt + R) = 108% (6 June–22 July) and the seasonal Q/(Snowmelt + R) = 99% (6 June–8 August). The 2012/13 ratios, while higher than those for a small wetland site (0.12 km^2) on Devon Island (0.39–0.50 – Rydén 1977) are typical for other polar desert catchments in the High Arctic (Young & Woo 2004b; Blaen et al. 2014 – Stuphallet cliffs). Runoff ratios for arctic watersheds can also be found greater than 100% owing to deep snowbeds in stream channels or on slopes lingering from the previous year (Woo 2012). At PBP, an estimate of 108% for the freshet runoff in 2013 is at the high end of typical runoff ratios, although still well within expected error limits of 15–20%, often attributed to arctic water budget studies (Young & Woo 2004b). Credibility of our water balance work also rests on the fact that Woo et al. (1983) and others have long argued that end-of-winter snow surveys in arctic landscapes provide a more
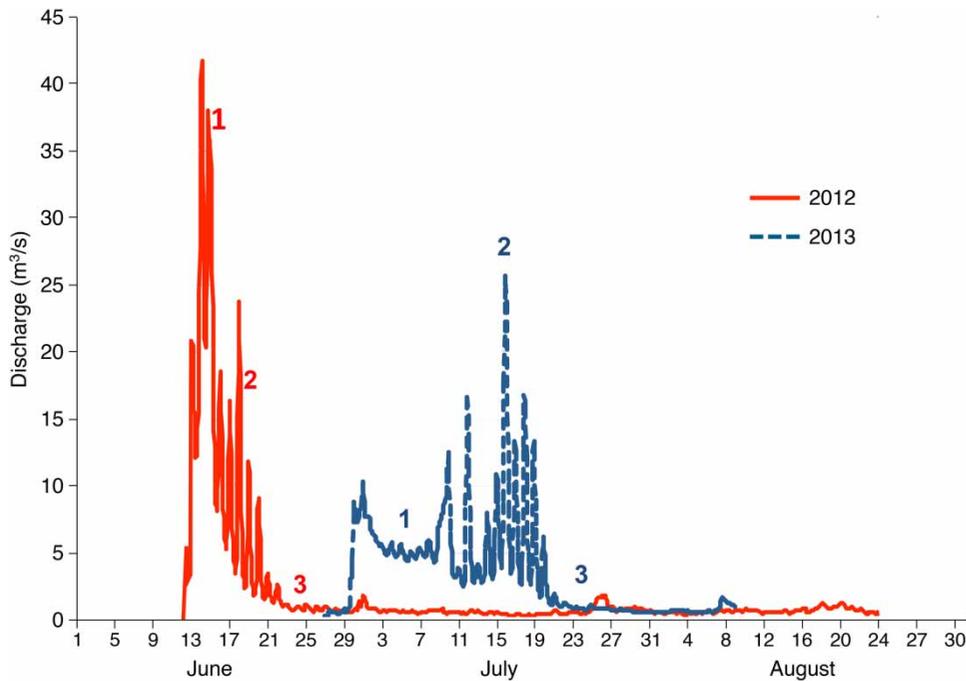
Figure 4 | Streamflow hydrograph for the eastern sector outlet, PBP in 2012, 2013. The influence of snowmelt from the northern half of the Pass is designated 1, the southern half snowmelt 2, and baseflow is 3.

reliable snow cover estimate than government snow gauge measurements, which routinely underestimate SWE (mm) in their surrounding catchments. Still, there are problems in our methodology which may have contributed to the higher runoff ratios, including that summer precipitation was not corrected for wind, which can enhance water budget errors up to 7% (Førland & Hanssen-Bauer 2000; Killingtveit et al. 2003). Second, deep snowpacks in incised valleys or hillslopes are often inadequately measured (Assini & Young 2012; Young et al. 2015). Assumptions concerning the distribution of the winter snowpack (SWE, mm) across terrain units may have also played a factor in the elevated runoff ratio in 2013. These two years were quite different in terms of weather conditions so the runoff patterns observed might provide a reliable indication of the upper and lower ranges possible at PBP. Clearly, additional years of monitoring would have been beneficial at PBP to better identify either seasonal trends or unambiguous patterns. This is unfortunately a limitation of short-term studies in remote arctic locations, a problem identified by many other arctic hydrologists (e.g., Young & Woo 2004a; Nowak & Hodson 2013; Blaen et al. 2014) and one which

will only worsen with the continual closure of northern monitoring stations (Laudon et al. 2017).

Comparison of streamflow to other Canadian High Arctic catchments

The early timing of streamflow and peak streamflow out of the PBP wetland (eastern sector) in 2012 was similar to nearby small hillslope catchments at PBP and to recent studies at Melville Island (East, West Rivers) (see Figures 5–7). Since 2003, these studies suggest an earlier start to runoff and peak response across the Canadian High Arctic islands. However, this pattern varied in 2013, where initiation of runoff was quite late and more akin to High Arctic streamflow studies in the 1970s and early 1980s – a much colder period than occurring today (see Woo & Young 2014). These strong swings in streamflow response from an exceptionally warm spring/summer to below average cold seasons is a reflection of the streamflow intensification process, where northern Canadian basins are responding to extreme shifts in climate (temperature) and precipitation (snow or rainfall receipt), a characteristic of
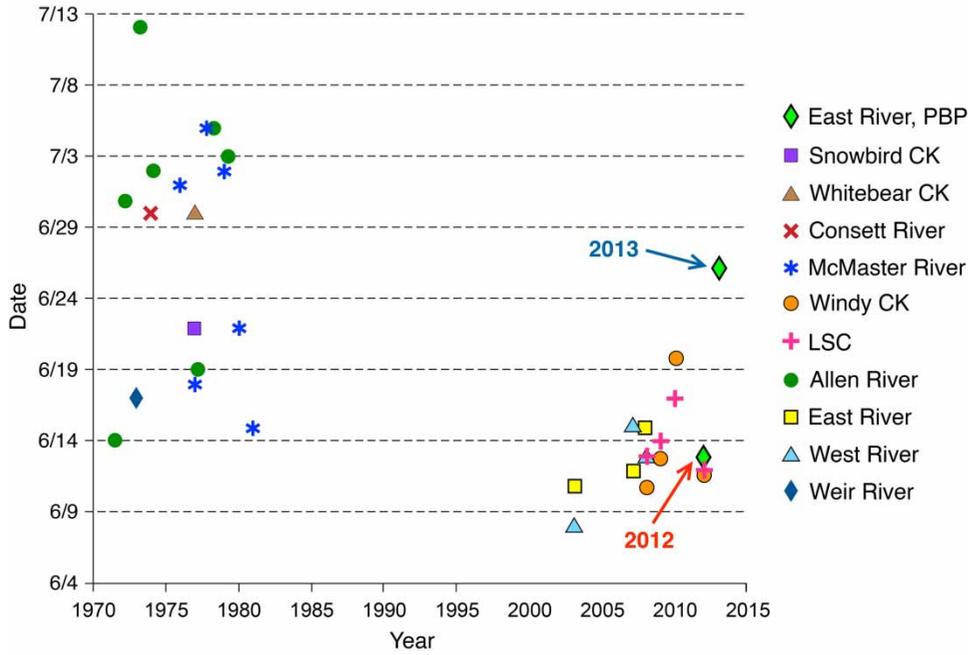
Figure 5 | First day of flow for PBP from the eastern sector watershed (East River, PBP) in comparison to other catchment and watershed studies. Diagram adapted after Young *et al.* (2015). Details about watersheds other than East River, PBP can be found in Young *et al.* (2015). 2003/2004 data for West and East Rivers, Cape Bounty are taken from Stewart & Lamoureux (2011). Arrows indicate first day of flow for the East River, PBP in 2012 and 2013.
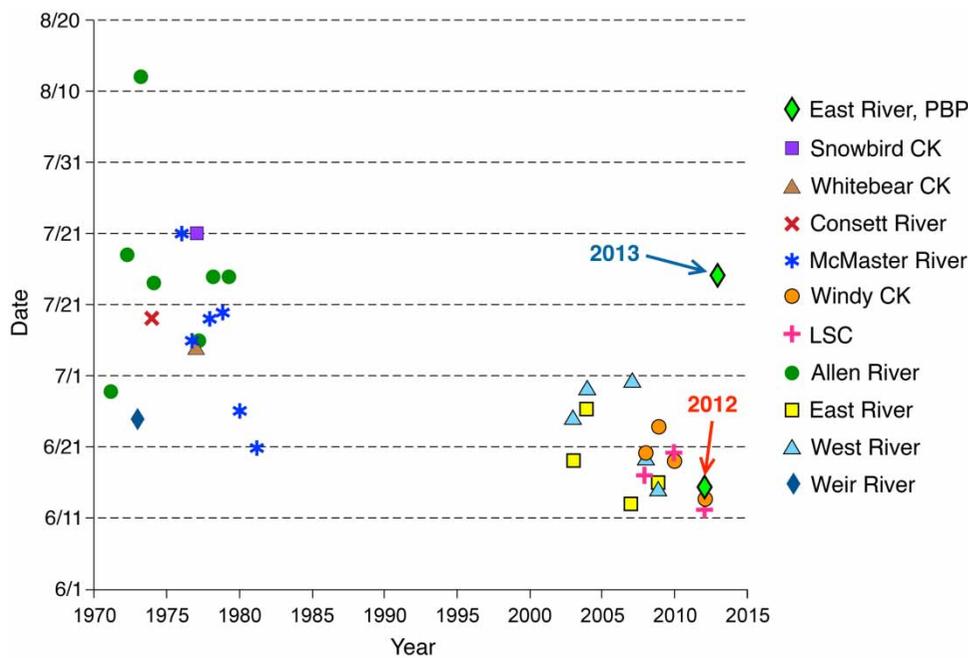

Figure 6 | Day of peak discharge from the eastern PBP outlet designated here as East River, PBP versus other arctic drainage basins. Details about watersheds other than East River, PBP can be found in Young *et al.* (2015). 2003/2004 data for West and East Rivers, Cape Bounty are taken from Stewart & Lamoureux (2011). Diagram adapted after Young *et al.* (2015). 2003/2004 data for West and East Rivers are taken from Stewart & Lamoureux (2011). Arrows indicate day of peak discharge for the East River, PBP in 2012 and 2013.
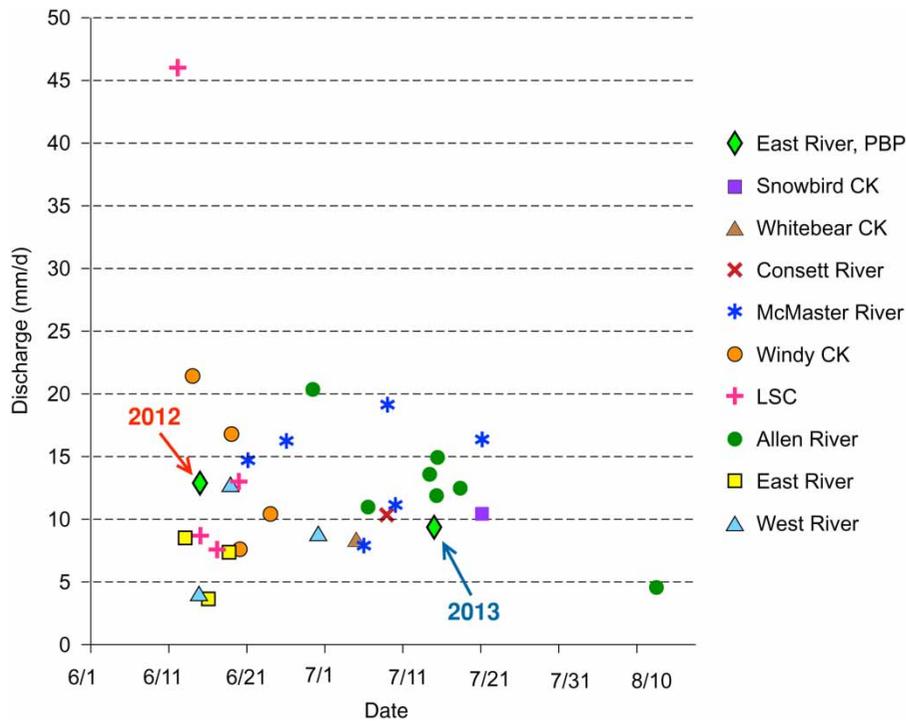
Figure 7 | Peak daily discharge (mm/d) at the eastern outlet (East River, PBP) versus other arctic drainage basins. Diagram adapted after Young et al. (2015). Details about watersheds other than East River, PBP can be found in Young et al. (2015). Arrows indicate peak daily discharge (mm/d) for the East River, PBP in 2012 and 2013.

the climate warming signal (Déry & Wood 2005; Déry et al. 2009). Several northern studies (e.g., Liljedahl et al. 2009; Kokelj & Jorgenson 2013; Nowak & Hodson 2013; Blaen et al. 2014; Karlsson et al. 2015) have also raised the importance of understanding the present status and shifting conditions of landscape micro-topography (% of high or low-centred polygons) and geomorphology (e.g., depth of active layer thaw, thermo-erosion processes, formation and decay of ground icings), as these factors can help modify rates of evaporation, water storage and runoff from arctic catchments. While gullying is prevalent on Bylot Island (Fortier & Allard 2007) and has contributed to rapid drainage and drying of a wetland area, similar rapid gully development is not yet apparent at PBP.

Seasonal water budgets

Given the differences in climate and snow cover in 2012 and 2013, it is reasonable to expect that the seasonal water budgets would vary (see Figure 8). In response to an early and warm spring, snowmelt in 2012 began earlier than in

2013. This was followed by a longer duration for runoff, evaporation and changes in wetland storage which aided in higher evaporation losses and greater wetland storage changes in 2012 than in 2013.

Compared to other arctic water budget studies (see Bolton et al. 2004; Young & Woo 2004b), evaporation losses can sometimes exceed summer precipitation inputs during warm seasons and this occurred in 2012 as well (E/R = 2.7). Here, soil moisture in the thawed active layer was the likely water source for continued evaporation (including transpiration) losses (Young et al. 2017). In 2013, evaporation losses were instead met by summer precipitation inputs as energy receipt was diminished (E/R = 0.97).

In 2012, there was about 24 mm of summer precipitation (from 24 to 31 July) which is not well reflected in the cumulative streamflow response (Figure 8). Our discharge records do indicate a significant rise in flow from 0.4 to 1.8 m^3/s on 24 July 24 in response to c. 5 mm of rainfall; however, this flow is still much lower than freshet levels (Figure 4). In addition, frost table measurements for 27 June
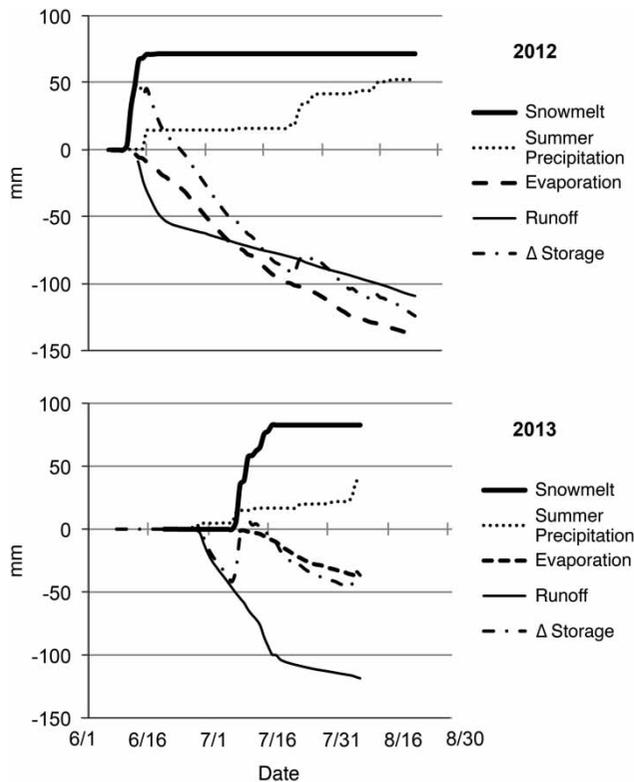
Figure 8 | Seasonal cumulative water budget of PBP eastern watershed sector: 2012 (5 June–23 August) top diagram and 2013 (6 June 20–8 August) bottom diagram. In 2012: $\sum$Snowmelt = +72 mm; $\sum$Summer precipitation = +52 mm; $\sum$Evaporation = −139 mm; $\sum$Runoff = −109 mm and $\sum\Delta$Storage = −124 mm. In 2013: $\sum$Snowmelt = +83 mm; $\sum$Summer precipitation = +38 mm; $\sum$Evaporation = −39 mm; $\sum$Runoff = −118 mm and $\sum\Delta$Storage = −37 mm. Note that in 2013, runoff (measured) begins prior to basin snowmelt (modelled). This indicates an error in the snowmelt model (see text). The shorter field season in 2013 is due to arctic logistics (a lack of helicopter support at the end of August).

indicate that ground thaw depths were already reaching upwards of −0.45 m (maximum depth) in the wet meadow and −0.51 m in polar desert uplands. Clearly, the active layer storage capacity in the basin was enhanced in 2012 due to a warm spring/summer (Table 1), and given the numerous ponds (see Figure 2(a)) and one large lake across this end of the Pass, it is surmised that much rainfall later in July could have been stored in the catchment with no hydrologic threshold being crossed. If it had been crossed, extensive overland flooding and a sizeable increase in streamflow would likely have occurred (see Bowling et al. 2003). We noted this same pattern for a small PBP hillslope creek in 2008. Runoff from the creek into the main wetland did not occur until the soil moisture storage in its upper catchment was satisfied by rainfall (Young et al. 2010).

Finally, the eastern sector of the PBP watershed that is being considered in this water budget study is quite large (estimated to be 102.6 km^2). Consequently, streamflow response to rainfall inputs at PBP over several days might not be as sensitive as small non-glacial catchments like the ones studied in Svalbard (<1 to 2 km^2) (see Blaen et al. 2014).

The storage term was large in 2012, reaching 124 mm, almost 3.4× greater than in 2013, and this can be attributed to 2013's cool summer and short field season in comparison to 2012's long and warm one (Table 1). Evaluation of the storage term, which is the residual in this study but also includes the error term, can be upwards of 20 to 25% for arctic basins (Young & Woo 2004a, 2004b). Finally, based on their research in Svalbard, a maritime High Arctic location, Nowak & Hodson (2013) recommend that when calculating the water balance for arctic catchments, special attention to precipitation in the shoulder seasons (e.g., September) should be considered, and early and late rainfall events should be assigned to the correct hydrological year. This could include winter rainfalls and the development of ground icings. While we have not yet observed winter rainfalls at PBP as air temperatures remain cold (Mudryk et al. 2017), ground icings can arise due to autumn rainfall on frozen ground and as such these phenomena should be evaluated in future water budget studies here.

CONCLUSIONS

Like other High Arctic basins, streamflow out of the eastern sector of PBP is dominated by seasonal snowmelt, which demonstrates a nival regime. In 2012, due to an early and quick snowmelt season, the first and highest runoff peaks responded to snowmelt from the northern half of the Pass. In contrast, during a cold and prolonged snowmelt season, initial runoff was fed by snowmelt from the northern half of the Pass, but it was not until snowmelt from the southern part reached the outlet that the highest peaks were reached in mid-July. In fact, streamflow response in 2013 was more akin to arctic catchments in the 1970s and 1980s, a period which was characterized by higher precipitation and colder conditions. Subsequently, the runoff ratios and water budgets differed between the years based on the

differences in precipitation and variations in climate. Given that future climate warming is expected to trigger extreme conditions (Stiegler *et al.* 2016) and the two years shown here (2012, 2013) provide a reasonable picture of this scenario, strong swings in streamflow out of large, low-gradient wetland watersheds in the High Arctic can be expected. Clearly, much research still needs to be carried out at PBP. Future studies here should examine the streamflow and water budget first from the western section of the wetland and then the entire watershed. If possible, precipitation inputs in the shoulder seasons, including the occurrence of ground icings, should be investigated to reveal their role in seasonal water budgets (Nowak & Hodson 2013). The Pass's location near Resolute, a government weather station and starting-off point for High Arctic expeditions makes it an ideal location for continued long-term hydrological studies.

Overseem & Syvitski (2010) indicate that the Arctic regions are undergoing rapid warming at rates much faster than the global average, and hydrological systems, especially rivers, are responding to these changes. Perhaps Buttle *et al.* (2012) state it best when they conclude that, 'there is a need to improve the monitoring of flow in such transient flow systems. We need to improve our ability to model temporary streamflow, particularly where controls involve snowmelt and permafrost. Temporary streams are critical but highly vulnerable hydrological and hydroecological systems that are often the first parts of the drainage network to respond to many natural and anthropogenic impacts'.

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