Recent multi-year streamflow regimes and water budgets of hillslope catchments in the Canadian High Arctic: evaluation and comparison to other small Arctic watershed studies

Kathy L. Young, Melissa J. Lafrenière, Scott F. Lamoureux, Anna Abnizova and Elizabeth A. Miller

ABSTRACT

This study evaluates whether a recent decline in snowcover extent for circumpolar regions is matched by changes in the seasonal streamflow regime of several small hillslope catchments on Bathurst and Melville Islands. This includes shifts in the timing of initiation, peak discharge and impacts on the spring–summer water budgets. Paired catchments (West and East) at the Cape Bounty Arctic Watershed Observatory (CBAWO) on Melville Island (74.9°N, 109.5°W) have been studied from the pre-snowmelt season to early August since 2003. They are low-rolling tundra catchments between 8.0 and 11.6 km^2 in area. Likewise, within the Polar Bear Pass (PBP) watershed, Bathurst Island (75.7°N 98.7°W), two hillslope basins, Windy Creek (4.2 km^2) and Landing Strip Creek (0.2 km^2) have been investigated since 2007. Detailed snow surveys were conducted each spring and streamflow estimates were made using the mid-section velocity method. Nival regimes continue to dominate in these basins but runoff ratios are variable between catchments, across islands, and from year-to-year. In comparison to earlier streamflow studies across the Queen Elizabeth Islands (QEIs), an earlier response to peak discharge and start of flow for these hillslope streams is confirmed. Water budgets for PBP, CBAWO differ from other small Arctic watersheds.

Key words | Canadian High Arctic, climate variability, snowcover, streamflow, water budgets, water resources

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INTRODUCTION

Like much of the Circumpolar Arctic, Canada's High Arctic islands are now experiencing a changing climate and evidence for these changes intersects across the terrestrial cryosphere. While warming in the Western Arctic has been ongoing for the last 20–30 years, and adjustments in streamflow are now appearing (Déry & Wood 2005; Déry *et al.* 2009), this warming did not start until about 1993 in the Eastern Canadian Arctic but so far has been continuing (Smith *et al.* 2010). Environmental manifestations of this warming include warming of permafrost and deepening of the active layer, loss of late-lying snowbeds (Woo & Young 2014), the drying, shrinkage and loss of ponds and lakes sometimes with a shift in ecological diversity (Smol & Douglas 2007a, 2007b), and a dramatic loss of glacial ice which is now considered irreversible (Lenaerts *et al.* 2013). Recently, Derksen & Brown (2012) have documented the rapid decline in snowcover extent across Pan-Arctic regions (April, May, June), with this reduction occurring most dramatically from 2005 onwards.

Snow is critical for High Arctic ecosystems; it replenishes ponds, lakes and desiccated wetlands. Meltwater from hillslope catchments deliver water, nutrients and sediments from upland terrain into these low-lying zones. Past studies in the Queen Elizabeth Islands (QEIs)

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(e.g., Wedel *et al.* 1977; McLaren 1981; Woo 1983; Woo 2000; Lewis *et al.* 2012, and others) have indicated streamflow regimes are dominated by nival regimes (snowmelt), with rainfall normally playing a secondary role in both streamflow generation and basin water budgets.

There is now a pressing need for reliable streamflow and water budget information in Arctic environments, especially from catchments with a range of scales and physiography (Spence *et al.* 2005). Spence *et al.* (2005) argue that new theories and modelled data cannot be tested if suitable data sets are not available. These researchers hypothesise that the sparse network of the Water Survey of Canada makes it difficult to predict streamflow in ungauged basins, particularly ones of interest to industry (Spence *et al.* 2005).

In terms of industry in High Arctic watersheds, details on the potential for flooding (timing, magnitude) are required for safe oil and gas pipeline installations across stream valleys, while mining extraction processes require secure water supplies too. Due to the high cost of diesel to generate electrical power, there is a growing desire to assess the potential of many Arctic streams for hydroelectric power, especially in Nunavut. Runoff also transfers nutrients and sediments into ponds, lakes, river and streams, and ultimately these waters empty into nearby Arctic coastal waters. To date, there is a particular interest in the delivery of carbon: dissolved inorganic and organic carbon into a range of water bodies and to evaluate its role in the global carbon cycle (Holmes *et al.* 2012; Lewis *et al.* 2012; Abnizova *et al.* 2014). Information on the timing and availability of water, including knowledge of hazards (both floods and droughts) are desired by local Inuit who have concerns about water supplies to their local water reservoirs, and the potential for destruction of infrastructure (bridges and sewage/water pipelines – e.g., Pangnirtung, Nunavut, June 2008).

Hydrological, paleological and biogeochemical studies were initiated around 2003 in small paired hillslope catchments at the Cape Bounty Arctic Watershed Observatory (CBAWO), Melville Island (74.9°N, 109.5°W). Similarly, since June 2007 at Polar Bear Pass (PBP), Bathurst Island (75.7°N 98.7°W), there has been a focus on both wetland and hillslope hydrology studies. In this particular study, we assess the impact of recent regional warming here (since 2005) (see Woo & Young 2014) and a reduction in snowcover extent as described by Derksen & Brown (2012) on both streamflow response and water budgets in these small hillslope catchments. We are particularly interested in exploring: (1) whether we are now seeing a shift in the streamflow regime, specifically, can we document a departure from a *Nival* to a *Mixed* or *Pluvial*-type of flow regime; and/or (2) if we are observing a decline in snowcover extent, is this translating into earlier streamflow, and an earlier and higher peak Q; and (3) in terms of seasonal water budgets, can we observe differences in the allocation of water resources between catchments, islands and years, and how do these estimates compare to earlier stream basin studies in this region and elsewhere (e.g., Alaskan Arctic, Northwest Canada, Eurasia)? Such basic knowledge on the water resources of this oft forgotten zone adjacent to the Northwest Passage is of critical importance to the environmental sustainability of this area, growing industrial services and the nearby Inuit community of Resolute Bay (Inuktitut: Qausuittuq). Moreover, it adds to the discussion concerning intensification of streamflow in Northern Canada (Déry *et al.* 2009) and elsewhere, and moves towards a broader understanding of the mechanisms accounting for such changes (Rawlins *et al.* 2010).

STUDY AREAS

Polar Bear Pass

PBP is located in the middle of Bathurst Island about 146 km northwest of Resolute Bay, Canada (Figure 1). The pass consists of a large, low-gradient wetland about 94 km^2 in area. It is made up of two large lakes, about 4,000 ponds, comprising 30% of the area, and wet meadow and patches of dry ground. It is bounded by low-rolling hills running east-west rising to 150–200 meters above sea level. These hills are notched by stream valleys (upwards of 50) which drain the upland areas of both snowmelt and rainwater (Young *et al.* 2013). Two typical hillslope stream catchments draining into the low-lying wetland were the focus of this study: Landing Strip Creek (LSC) – 0.2 km^2, and Windy Creek (WC) – 4.2 km^2 (Figure 2). PBP is considered a biological oasis although a comparison of its climate with Resolute Bay (Young & Labine 2010) suggests

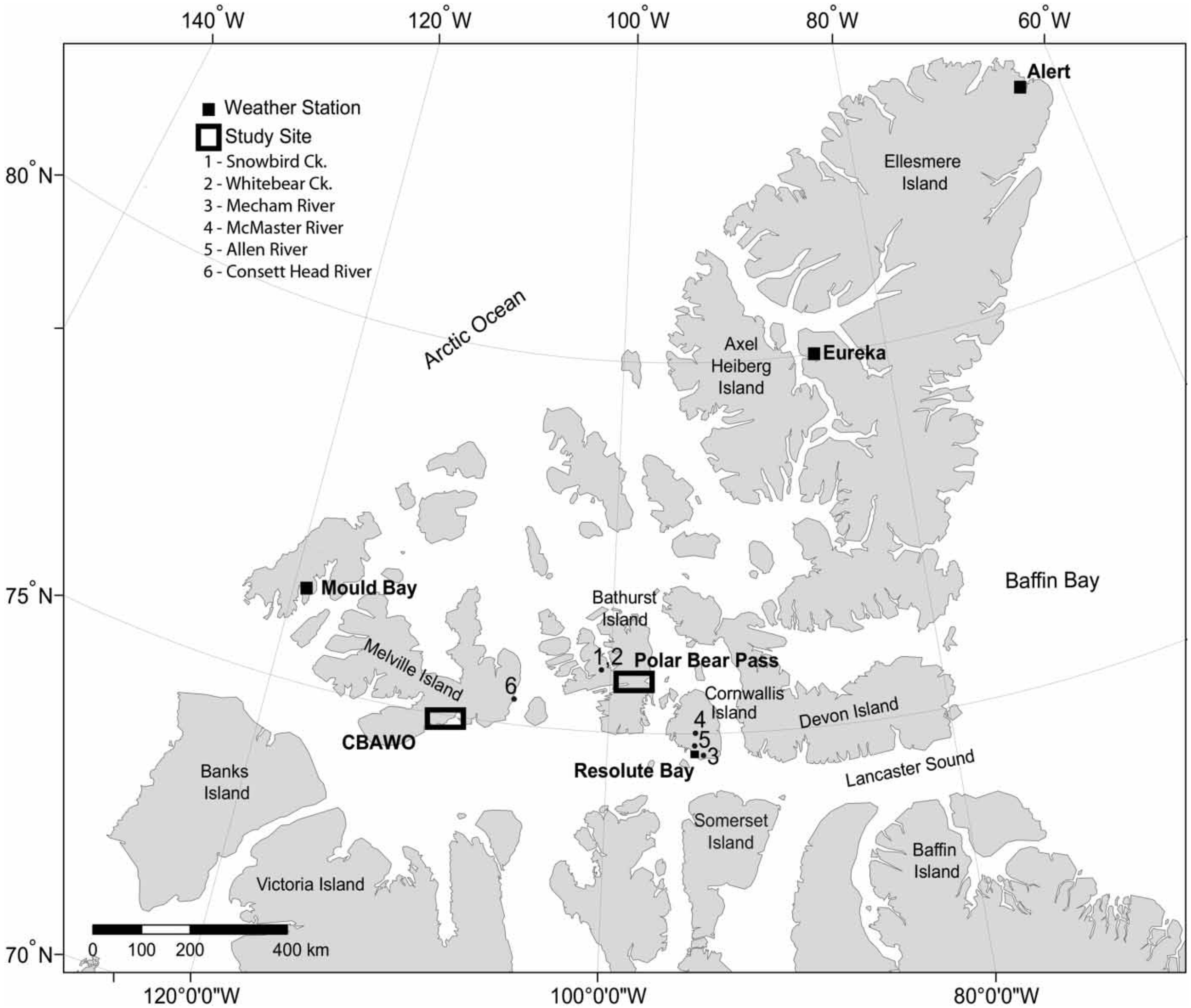

Figure 1 | Map of the QEIs, the northernmost cluster of islands in the Canadian Archipelago above Lancaster Sound, showing the location of the main study sites of PBP, Bathurst Island and CBAWO, Melville Island. Locations of the main government weather stations and previous regional watershed studies: Snowbird Creek (1); Whitebear Creek (2); Mecham River (3); McMaster River (4); Allen River (5) and Consett Head River (6) are also indicated. More details about these older catchments can be found in Table 1. Note that the Mould Bay government weather station now consists of an AWS. It is located about 336 km northwest of Cape Bounty (CBAWO).

that it can also be considered to have a polar desert climate: long, cold winters and short, cool summers. The region is underlain by continuous permafrost and active layers average 0.3–0.5 m in boggy, wet soils, while dry polar desert ridges can reach upwards of 1 m in warmer years (e.g., 2010, 2012). LSC (unofficial name) is a first order creek, typical of others in the Pass (see Figures 2(a) and 2(c)). It has a south-facing slope, and is sparsely vegetated, comprising polar desert soils (porosity = 0.50, organic content <4%). The upper part of the catchment is bowl-shaped and a lingering late-lying snowbed forms along the upper ridge. Ground thaw here can reach 1.0 m. The main channel is narrow and steep and waters drain into a wet meadow with small tundra ponds. The upper soils here are organic and have a porosity of 0.93 (Young *et al.* 2010). WC (unofficial name) is the larger catchment (4.2 km^2) and has three main tributaries (Figures 2(b) and 2(d)). The east-west tributary (Moss Creek, unofficial name) has an extensive organic layer 10–160 mm in thickness over gravelly ground and a late-lying snowbed typically lies along the north-facing slope. The main channel of WC branches above Moss Creek with one tending northeast to southwest, while the other is northwest to southeast facing. The rocky channel length is 2.73 km and the average slope is 3.9°. Polar

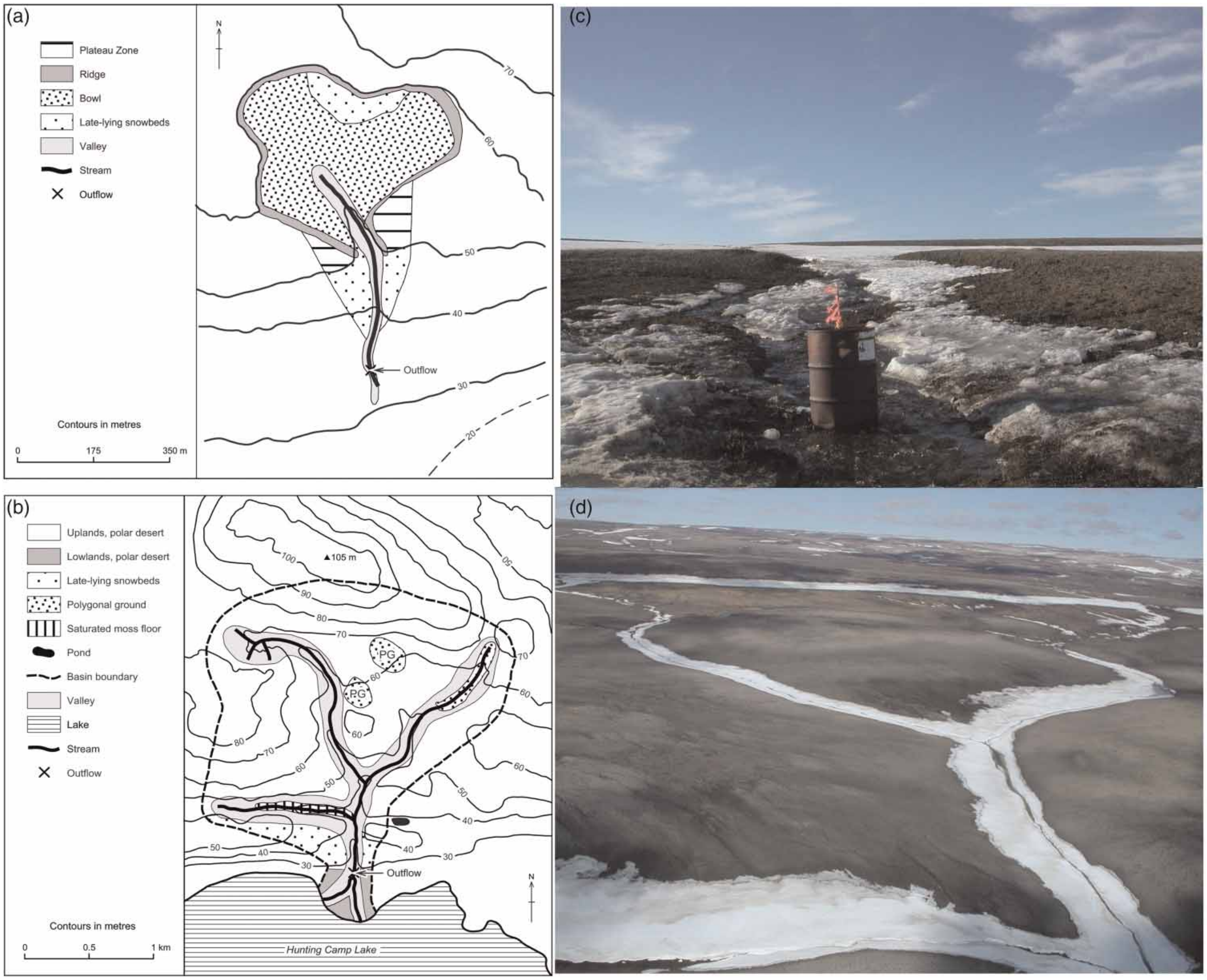

Figure 2 | Topographical maps of the hillslope catchments at PBP: (a) LSC and (b) WC. Diagrams are adapted from Young *et al.* (2010). Also shown are photographs of the basin at (c) LSC outlet looking upslope and northward, June 15, 2012 and (d) aerial view of the upper portion of WC, June 18, 2012, looking north. Note that AWSs were located outside of the catchments.

desert soils comprise most of the basin although pockets of ice-rich polygonal terrain with shallow frost tables occur here (Young *et al.* 2010).

Cape Bounty Arctic Watershed Observatory

CBAWO is located on the south-central coast of Melville Island (Figure 1). Paired basins have been the focus of study since 2003 (Figure 3). Similar to PBP, these catchments are underlain by continuous permafrost and have a maximum active layer of less than 1 m. The sparsely vegetated catchments are composed of low-relief hills (<120 meters above sea level), and prostrate tundra vegetation is typical on the uplands while wet sedge meadows occur in low-lying wetter areas. The area can be considered to have a polar desert climate, with annual precipitation approaching 111 mm and July temperatures of 4 °C (1971–2000) at Mould Bay (76.3°N, 119.5°W), the nearest government weather station, within 336 km of CBAWO (Stewart & Lamoureux 2011). Both catchments are south-facing and have tributaries which feed a main channel draining into lakes but the West River (unofficial name) catchment

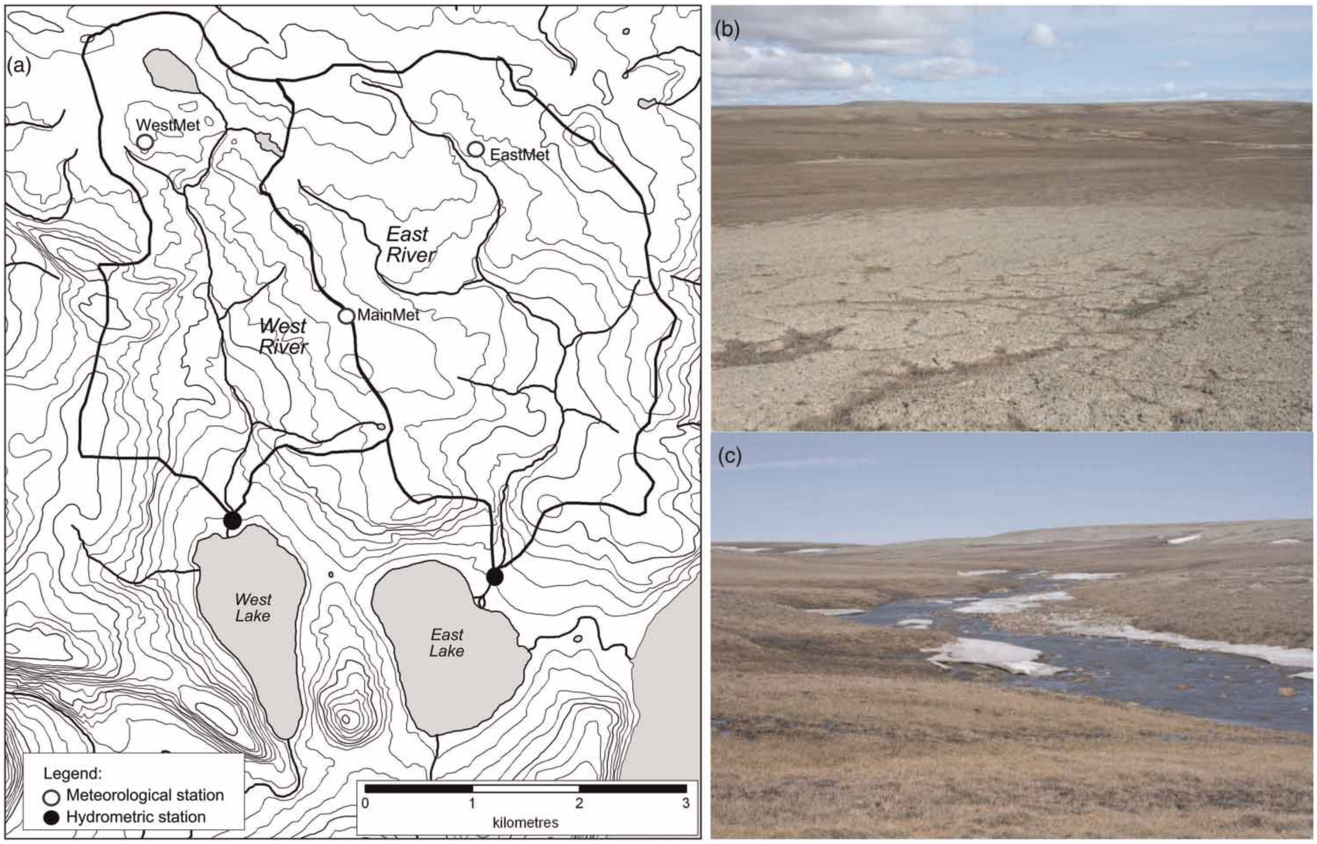

Figure 3 | Topographical map showing the hillslope catchments (East and West River basins) at the Cape Bounty Watershed Observatory (CBAWO) (a). Also shown are photographs of East River basin looking north (b), and West River catchment, looking northwest (c). The source for the topographic map is NTS map 78F/15. Note that AWSs were located in the basins.

($8.0\ km^2$) has a steeper gradient and deeper gullies and channels than the East River (unofficial name) catchment ($11.6\ km^2$). This usually results in more snow for the western basin (Stewart & Lamoureux 2011).

METHODOLOGY

The theoretical framework for this research is the water budget, which evaluates for a given period the amount of water gained, lost and stored in a defined catchment (Young & Woo 2004a)

$$P(Sn + R) - E - Q = \Delta S + \xi \quad (1)$$

where P is precipitation input including snowfall (Sn), specifically the end-of-winter snowpack, in snow water equivalent terms (Kane *et al.* 2004) and R, rainfall which sometimes includes summer snowfall, E is evaporation, which includes transpiration and Q is runoff. The term ΔS is change in storage (including snow storage in lingering snowpacks, soil moisture, ground ice and groundwater storage) and ξ is the error term. During the spring melt period, snow (Sn) will eventually ablate (melt, sublimate), or it may enter into storage (Young & Woo 2004a; Woo & Young 2014). In this study, during the spring melt and summer thaw period, Sn in Equation (1) is replaced by M (snowmelt) (see Marsh *et al.* 2004).

In seasonal catchment water budgets, rarely is ξ estimated separately from ΔS (see Kane & Yang 2004; Marsh *et al.* 2004; Young & Woo 2004a, b) and this is also the case for this study, i.e., all errors related to gauge measurements, analytical and human errors can potentially reside on the right side of the equation in ($\Delta S + \xi$). Woo (2012,

p. 504) remarks that due to limited observations in remote locations and measurement problems 'it is not easy to close the water balance'. He further suggests that 'some level of uncertainty has to be accepted but within specified limits of tolerance, the results should enable qualitative assessment of the hydrologic status of the basin'.

Polar Bear Pass

At PBP end-of-winter snow surveys (late May) were conducted to assess the snow amount P(Sn) in snow water equivalent terms (SWE, mm) in each one of these basins. Methodology followed after Woo (1997) and consisted of a series of transects across varying terrain types making up the catchments. Snow depth was typically made every 10 to 50 m along long transects (plateau) and every 2 to 5 m across shorter ones (valley bottoms, ponds). Except for ponds, each snow depth measurement was typically the average of four measurements. Snow density was measured using a Meteorological Service of Canada snow tube at the start and end of each transect, with more frequent sampling along long transects >1 km. Daily snowmelt (M) for the terrain types in the catchments (plateau, valley, slope) was estimated using a surface energy budget snowmelt model (Woo & Young 2004) which was previously deemed applicable to this site (Young et al. 2010; Assini & Young 2012; Young et al. 2013), and elsewhere (Woo & Young 2004; Young 2008). Meteorological inputs to the model (air temperature, relative humidity, solar radiation, precipitation, wind speed and pressure) were available at an automatic weather station (AWS) centrally located in the wetland (referred to here as CAWS) and were complemented by climate data from the Plateau AWS. Both of these AWSs are within 1 km of each other, and about the same distance from each study site. More details about the surface energy-budget melt model used in this study can be found in both Woo & Young (2004) and Young et al. (2010). Snowmelt estimates were compared with direct measurements of surface snow ablation following after Heron & Woo (1978) at terrain types typical of the catchments (pond, plateau, late-lying snowbed, wet meadow). Here a marked line (~2 m long) was held taut by two dowels placed in the snowpack. Ten measurements from the middle of the line to the snowpack at intervals of 10 cm were made daily and then averaged. These snow lowering measurements together with five snow density measurements obtained from the near surface (each snow sample being 200 cm^3) allowed the surface snow water equivalent to be determined (mm). In mid-June snow cover distribution in the catchments was validated with low-level aerial photography via a helicopter.

Summer precipitation (R-rainfall and snow, mm) areally weighted for each catchment was quantified with a recording precipitation gauge at the AWSs. This was supplemented by manual raingauges at each catchment (see Young et al. (2010) for more details). The Priestley–Taylor (P–T) approach (Priestley & Taylor 1972) was used to assess evaporation (E) from the wet meadow and saturated ground following snowmelt using $\alpha = 1.26$. For moist and drying ground, the α term was adjusted to near-surface soil moisture conditions (0–15 cm) using the algorithm developed by Marsh et al. (1981) for polar desert terrain. Near surface volumetric soil moisture (θ) was determined in both wet meadows and polar desert using either continuously measuring Campbell Scientific TDR probes or Echo probes wired into Campbell Scientific data-loggers. These hourly estimates were confirmed by regular volumetric soil moisture measurements with a hand-held Theta probe (see Young et al. (2010) for more details). Occasionally (e.g., 2008), environmental data were missing for the polar desert terrain site, either net radiation (Q^*), ground heat flux (Qg) or volumetric soil moisture (θ). At these times, evaporation was quantified using the empirical approach described by Stewart & Rouse (1976). This method calls for inputs of solar radiation and air temperature, and was originally developed for Subarctic areas. Undoubtedly, the use of this alternative method likely led to some error in evaporation estimates for polar desert ground. Others have reported on the difficulty of estimating evaporation including transpiration in northern latitudes (e.g., Vasilenko 2004; Liljedahl et al. 2011). Like snowmelt, evaporation was areally weighted for each catchment based on terrain type, and climate data for evaporation estimates came from both the Plateau AWS and the CAWS (see Young & Labine (2010) and Young et al. (2010) for additional details).

Seasonal runoff from each catchment was captured near its exit and standard techniques were used to make estimates of this discharge. Stage was monitored by recording water level sensors with Hobo pressure transducers (±3.0 mm) or

Ecotones (±25.4 mm). Frequent direct measurements of stilling well stages were made with a metric ruler (±5.0 mm). They provided an additional check on the reliability of continuous stage measurements at different time intervals, and/or were used to correct stage when values drifted. Direct discharge measurements at both low and high flows were made to develop reliable stage-discharge 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. These equations with an R^2 typically >0.80 were then used to adjust the water stage into a continuous discharge record. Streamflow estimates can amount to 10 to 15% error on average (Young *et al.* 2010). Further details on the current metering equipment deployed at each site can be found in Young *et al.* (2010). Determination of terrain areas in each catchment (±5%) is described in Young *et al.* (2010). As mentioned above, these areas were used to areally weight daily snow, evaporation and rainfall for both LSC and WC basins. Exceedance probability diagrams were produced of both snowmelt and post-snowmelt streamflow, following after Woo (2000), in order to assess whether departures from a nival stream regime were now occurring in these stream basins. Here, hydrograph separation between these periods (snowmelt vs. post-snowmelt) was made subjectively by extending the recession limb of the last major snowmelt pulse to the x-axis marking time (Ward & Robinson 2000). Simple non-parametric tests (Mann–Whitney) were run to establish differences in the start of flow, timing of peak flow and peak runoff amounts from LSC, WC and CBAWO streams (here referred to as 2000 group) compared to earlier studies (here referred to as 1970s group).

The right side of Equation (1) is the residual in this study ($\Delta S + \xi$), and as mentioned earlier, all of the errors can reside in this term(s). Storage can include snow storage in lingering snowpacks, soil moisture, ground ice and groundwater storage. Like previous studies here and elsewhere, uncertainty in our basin assessments likely falls between 15 and 20% given the remote location of this study, and limitations with snowmelt, streamflow and evaporation estimates. In particular, any errors in terrain unit areas will be reflected in precipitation, snowmelt and evaporation values. For instance, it is well known that water can cross boundaries both on the surface and below the ground due to a variable snowcover distribution and an evolving frost table. Melt-out of unaccounted late-lying snow or ground ice can lead to overestimates in the runoff ratio (Q/P) (Young & Woo 2004a, b; Woo 2012; Woo & Young 2014).

Cape Bounty Arctic Watershed Observatory

Similar to PBP, a terrain-based snow survey was routinely conducted at CBAWO in late May or early June prior to melt. The methodology differs slightly than the one developed for PBP. Here, 21 snow survey transects were set up across the basins. Each transect was comprised of 10 depth measurements and three snow density measurements. Mean SWE (mm) was estimated for each transect and then spatially averaged for each catchment based on terrain type (channel, slope and plateau) (Lamoureux *et al.* 2006). At the meteorological stations ($n = 3$), air temperature was measured at a 1.5-m height with Onset Hobo H8 loggers (±0.4 °C) or a Humirel HTM2500 sensor logged with an Unidata Prologger at MainMet (±0.2 °C), with precipitation measured with a Davies tipping bucket gauge (0.2 mm). Incoming solar radiation at MainMet was monitored with a Davies industrial sensor (5%) and soil moisture with Decagon Echo 10 cm probes (5%) and Onset U12 loggers. Hydrometric stations were located near the outlet of both East and West Rivers. Stage was recorded with Onset U20 water level loggers (±2 mm) at 10-minute intervals. Stream velocity was measured with a Swoffer impellor meter usually every second day during the season and rating curves (stage versus discharge) for the catchment routinely exceeded an R^2 of 0.80 (Lewis *et al.* 2012). In the absence of ground heat flux data (Qg) and net radiation (Q^*) data for some years, evaporation was estimated using the empirical approach by Stewart & Rouse (1976). The previous study by Young *et al.* (2010) showed that it was applicable for polar desert terrain.

RESULTS AND DISCUSSION

Type of streamflow regime

Figure 4 provides details on the recent seasonal streamflow regimes from PBP and CBAWO, and the exceedance probabilities of streamflow during the snowmelt and post-snowmelt

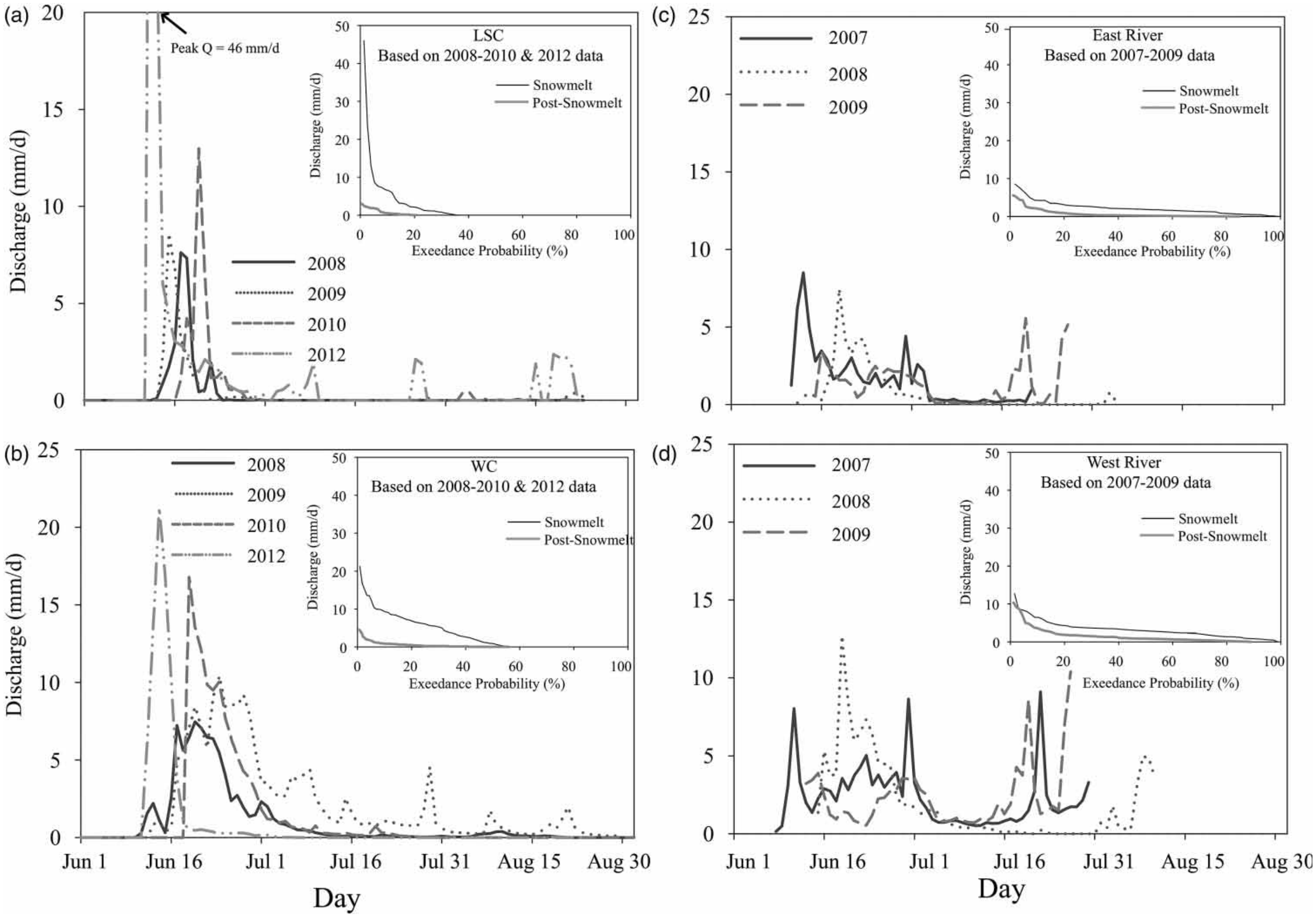

Figure 4 | Seasonal streamflow regimes (mm/d) at PBP: (a) LSC and (b) WC; and at CBAWO: (c) East River and (d) West River. Inset diagrams show the exceedance probability of streamflow for the snowmelt and post-snowmelt periods for each catchment. Peak discharge at LSC in 2012 was estimated to be 46 mm/d.

periods. Clearly, a nival regime still exists for these hillslope streams where the bulk of the runoff is generated by snowmelt. Peak streamflow was produced by snowmelt at PBP with rainfall in the post-snowmelt period never leading to runoff exceeding 5 mm/d (see inset diagram). This streamflow pattern differed for CBAWO, where in 2007 and 2009, the greatest peaks were generated by rainfall with the West River exhibiting higher peaks than the East River. Kane *et al.* (2009) indicated that in the Alaskan Arctic, for small and medium-sized basins (<200 km^2) floods of record are usually caused by rainfall rather than snowmelt. Other studies in the High Arctic confirm the importance of summer rains in generating peak flows (e.g., Cogley & McCann 1976; Woo & Guan 2006). Figure 5 indicates that climatic conditions, including summer precipitation totals (aside from 2007) are comparable for PBP and CBAWO, which suggests that differences in their runoff patterns may be linked to local site conditions (i.e., basin topography, soils, ground ice).

Day of peak snowmelt Q

Day of peak discharge generated from snowmelt for the study sites (Figure 6) is compared with previous streamflow studies across Bathurst, Melville and Cornwallis Islands (prior to 1985), and available Water Survey of Canada data from Cornwallis Island (i.e., the Allen and Mecham Rivers; see Table 1 for further details on these sites). Generally, peak discharges at PBP and CBAWO occur before June 29. In comparison, catchment studies in the 1970s/1980s suggest that discharge peaked after July 4 although there are occurrences where peak discharge occurred much earlier (e.g., McMaster and Mecham Rivers). A non-parametric Mann–Whitney test

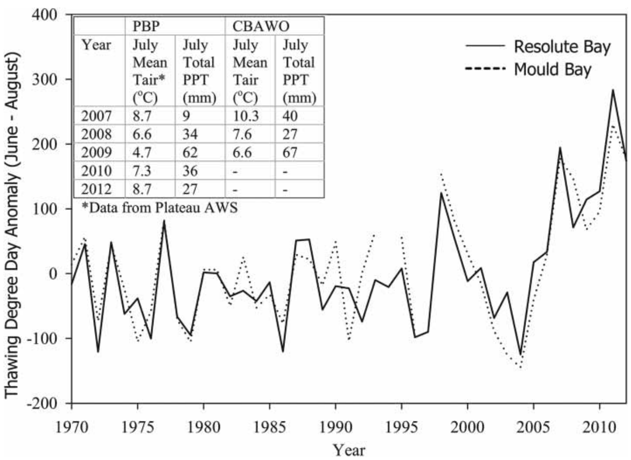

Figure 5 | Thawing degree day anomaly (June to August) for Resolute Bay (solid line) and Mould Bay (dashed line) weather stations plotted from 1970 onwards. Note the steep warming since 2005. The term *Thawing Degree Day* refers to the sum of daily air temperatures >0°C for June–August for each year. Thawing degree day anomaly values plotted here, show how yearly estimates relate to the long-term average of 257 thawing degree days for Resolute Bay (1953–2012), and 238 thawing degree days for Mould Bay (1953–2012). Inset table shows July average air temperature and precipitation totals for PBP and CBAWO.

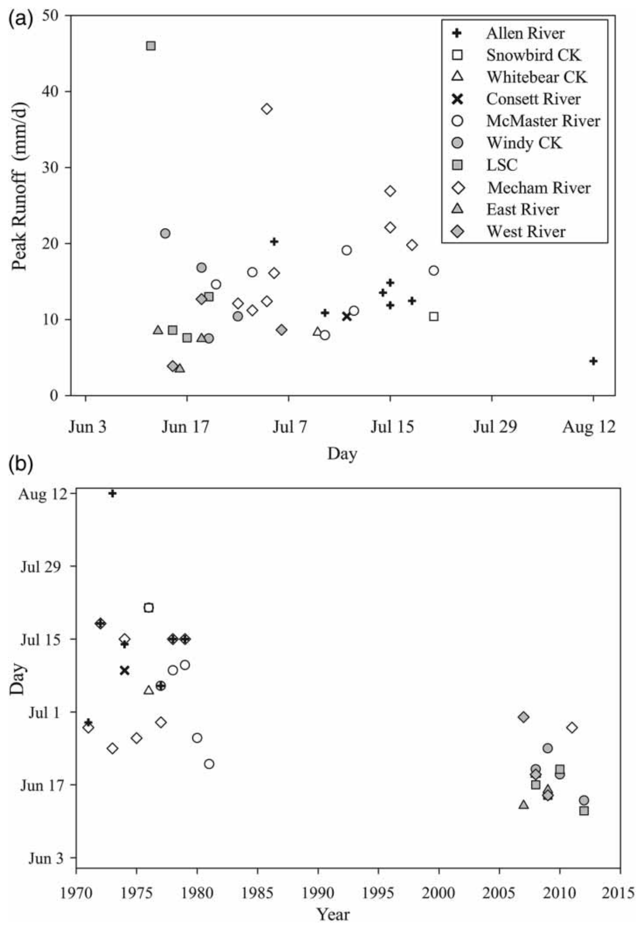

Figure 6 | Comparison of peak runoff from the study sites (PBP, CBAWO) in relation to past streamflow studies across the QEIs (Melville, Bathurst and Cornwallis Islands) and Water Survey of Canada data from Cornwallis Island. Diagrams show (a) peak runoff (mm/d) in relation to time (calendar days) for sites and (b) calendar day of peak runoff in relation to the year of record. Please note that in 2011 Water Survey of Canada resumed flow measurements at the Mecham River, Cornwallis Island. Solid symbols were used for this study and open symbols for earlier studies, including Mecham River in 2011.

concludes that the median date for peak flow is significantly different between the earlier and later streamflow studies ($\rho < 0.05$). This indicates that peak snowmelt discharges are occurring earlier across Bathurst and Melville Islands, athough these data should be viewed with some caution given the limited years of study. Aside from LSC, the relative magnitude of peak discharges at PBP and CBAWO sites appear to be similar to previous studies, though a Mann–Whitney non-parametric test (median peak flow) confirms that the two groups are quite different ($\rho < 0.05$). An extreme runoff value for LSC in 2012 arises from its small size ($0.2\ km^2$) and the rapid drainage of its channel after a slush flow was breached. Others have reported on the flashy response in hydrographs due to small catchment size (Kane *et al.* 2008) and/or occurrence of slush flows (e.g., Braun *et al.* 2000). Variability in flow regimes is also considered an example of the recent intensification of the hydrological cycle in Northern Canada (mainland) (Déry *et al.* 2009) and elsewhere (Rawlins *et al.* 2010).

Onset of snowmelt runoff

For most years, the initiation of snowmelt runoff for our study catchments is earlier than previous studies cited in Figure 7 and Table 1, or described elsewhere (McCann *et al.* 1972). This is likely in response to the recent summer climate warming across the Canadian High Arctic (see Figure 5), a finding confirmed by others conducting cryospheric research elsewhere in the Canadian High Arctic (Fisher *et al.* 2012; Zdanowicz *et al.* 2012; Gascon *et al.* 2013; Lenaerts *et al.* 2013; Woo & Young 2014). This difference in the timing of snowmelt runoff (median date) is also confirmed by a Mann–Whitney non-parametric test ($\rho < 0.05$). A regional trend towards an earlier start to snowmelt runoff has been observed in Eurasian Arctic rivers and Tan *et al.* (2011) attribute it to an upward trend in air temperatures.

Table 1 | Additional information on past streamflow studies located across the QEIs, Canadian High Arctic and mentioned in the text. Relevant hydrographic data are also plotted in Figures 6 and 7

Basin	Description and area (km^2)	Location (Lat., Long.)	Years of record	Source
Snowbird Creek, Western Bathurst Island	Polar desert, ridged upland to lowland, barren, only 10% vegetated, sandy, gravelly soils (61.1 km^2 above gauge)	76°N, 101°W	1976	Wedel *et al.* (1977)
Whitebear Creek, sub-catchment of Snowbird Creek, Western Bathurst Island	Polar desert, low-lying uplands/lowlands; barren, with vegetation concentrated in valley bottoms, sandy, gravelly soils (8.5 km^2)	76°N, 101°W	1976	Wedel *et al.* (1977)
Mecham River, Southern Cornwallis Island	Polar desert, barren ground and bedrock with sparse vegetation (1–5%) (McCann *et al.* 1972) (86.8 km^2)	74.7°N, 94.8°W	1971–1975, 1977–1979, 2011	Water Survey of Canada (www.wsc.ec.gc.ca)
McMaster River, Southwestern Cornwallis Island	Polar desert, barren ground and bedrock (33 km^2)	75°N, 95°W	1976–1981	Streamflow data courtesy of M. K. Woo, Professor Emeritus, McMaster University
Allen River, Southwestern Cornwallis Island	Polar desert, barren ground and bedrock (448 km^2)	74.8°N, 95.1°W	1971–1974, 1977–1979	Water Survey of Canada (www.wsc.ec.gc.ca)
Consett Head River, Eastern Melville Island	Polar desert, barren, low-lying hills, sandy, gravelly soils (117 km^2)	75.4°N, 105°W	1974	McLaren (1981)

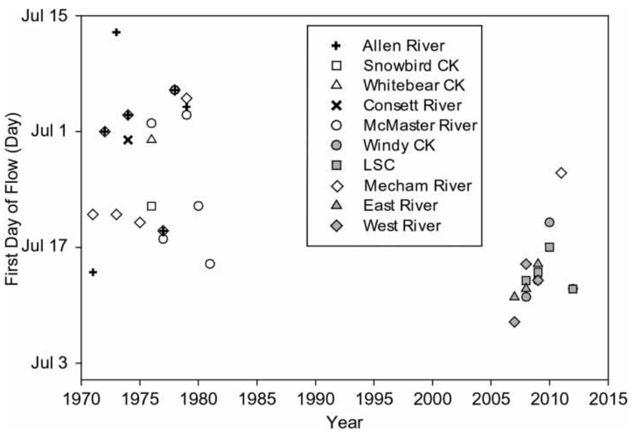

Figure 7 | Initiation of runoff versus year for the study sites (PBP, CBAWO) in relation to past streamflow studies across the QEIs (Melville, Bathurst and Cornwallis Islands). Please note that in 2011 Water Survey of Canada resumed flow measurements at the Mecham River, Cornwallis Island. Solid symbols were used for this study and open symbols for earlier studies, including Mecham River in 2011.

Seasonal water budgets (PBP and CBAWO)

Figure 8 illustrates the seasonal water budgets for WC and LSC (PBP) (Figures 8(a) and 8(b)) from 2008 to 2010, 2012 and the East and West catchments of CBAWO from 2007 to 2009 (Figures 8(c) and 8(d). Runoff ratios (Q/P – %) are also shown in the diagrams.

At the study hillslope creeks at PBP there was more rainfall than snowmelt in 2008 and 2009. Streamflow was variable from year-to-year with higher flows being maintained in 2009, a cool and rainy season. The residual term ($\Delta S + \xi$) differed between the catchments and remained variable. Negative storage persisted at WC over the years, implying that water is being taken out of storage and contributes to runoff. This might be accommodated by snowmelt from residual snowbeds, either lying in the lee

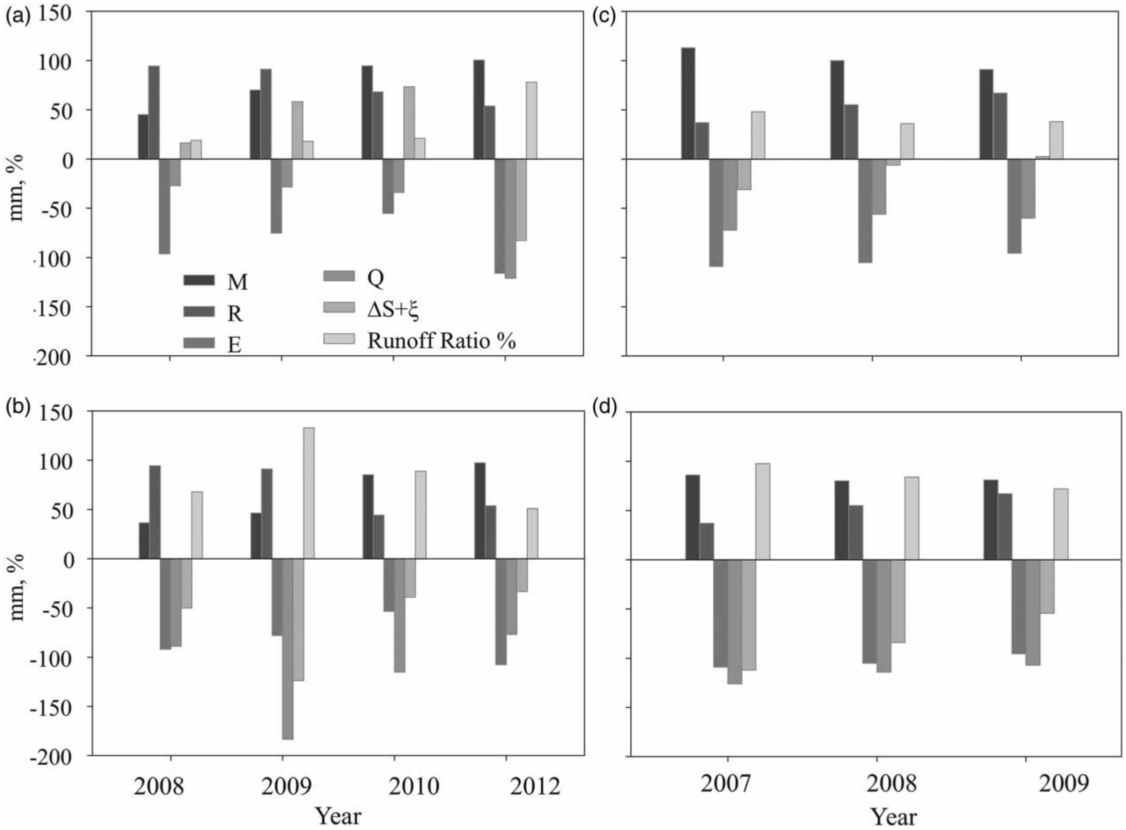

Figure 8 | Seasonal water budgets for PBP (2008–2010, 2012): (a) LSC: 2008 (May 31–August 20), 2009 (May 31–August 23), 2010 (May 31–August 3), 2012 (June 6–August 24); (b) WC and CBAWO (2007–2009); (c) East River: 2007 (June 11–July 21), 2008 (June 14–August 10) and 2009 (July 16–July 28); and (d) West River: 2007 (June 10–July 24), 2008 (June 14–August 10) and 2009 (June 12–July 28). Also shown are the seasonal runoff ratios (Q/P – %). Here, precipitation inputs are defined as snowmelt (M) and rainfall (R). Water losses include evaporation (including transpiration) (E), runoff (Q), while change in basin storage is ($\Delta S + \xi$), as defined in Equation (1).

of slopes or in the stream valleys. Ground ice melt might also supplement flow. The presence of shallow frost tables in ice-rich zones (vegetated polygons) can limit infiltration helping to shed rainfall into the stream channel (Young *et al.* 2010). Runoff ratios are generally less than 1 (100%), except for in 2009.

At CBAWO, more snowmelt than rainfall is accounted for in the catchments. Evaporation losses exceeded runoff in the East River but not the West River. The East River meanders more than the West River (i.e., shows a higher rate of curvature – see Figure 3(a)). It also has several more east-west tributaries allowing for a higher proportion of south-facing slopes in the catchment. Subsurface storage potential is likely greater here too as it retains a deeper active layer than the West River basin (McLeod 2008; MSc thesis, Department of Geography, Queen's University, Kingston, Ontario 2008, unpublished). These two landscape factors in the East River basin: south-facing slopes, deep storage potential for soil moisture would likely enhance evaporative losses. Storage was negative in both catchments but levels were slightly more negative in the West River. Runoff ratios in both catchments were less than 100% with the East River never exceeding 50%.

Daily water balance at PBP

Figure 9 provides an example of the daily water balance for LSC and WC at PBP (2009, 2010 and 2012). No fieldwork (spring/summer) was conducted in PBP in 2011 due to medical reasons. The 2008 daily water balance data for these streams has been previously reported so is not shown here (see Young *et al.* 2010). For both catchments, inputs of snowmelt into the catchment and losses from discharge dominate the early part of the spring season. The

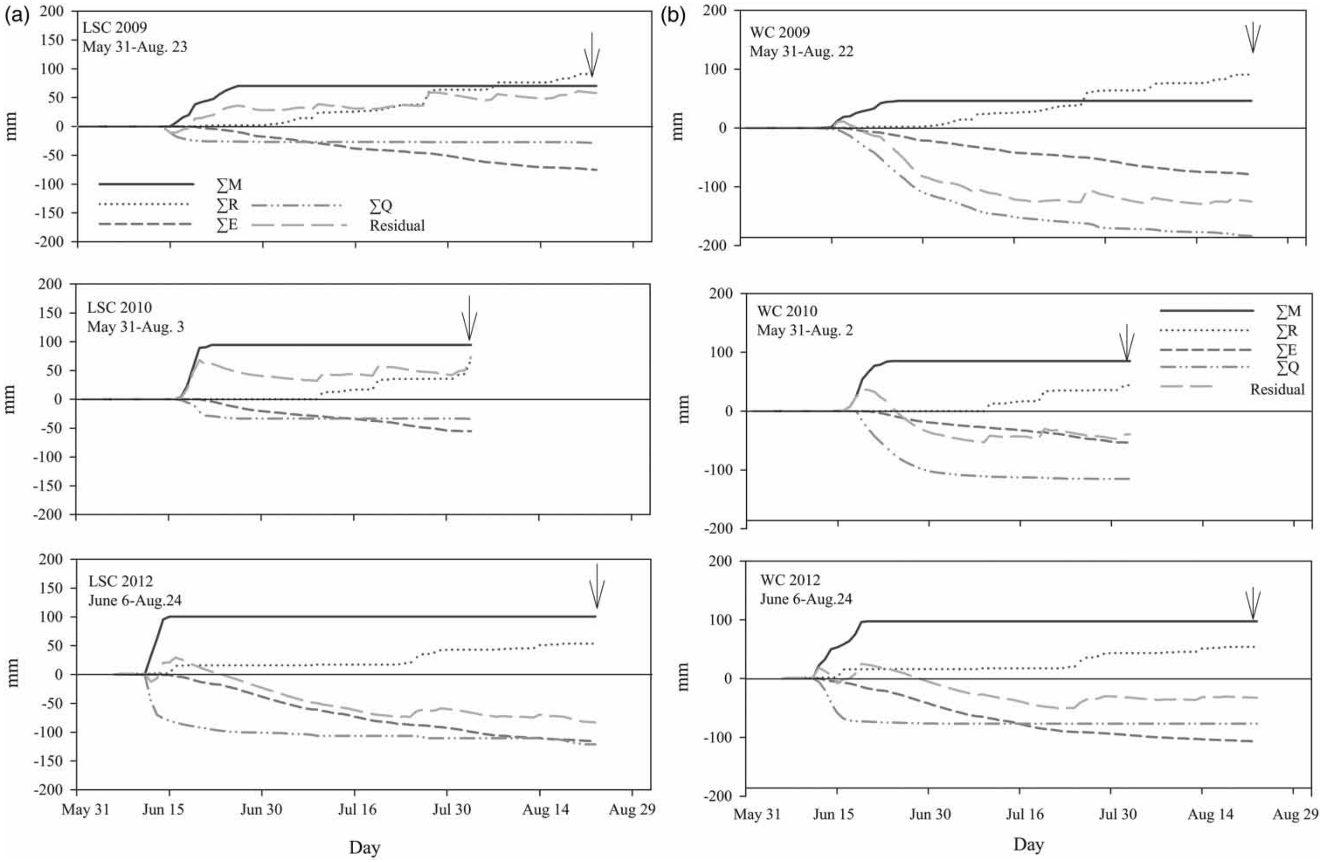

Figure 9 | Daily cumulative water balance values for (a) LSC and (b) WC at PBP for the spring, summer seasons of 2009, 2010 and 2012. Note no water basin study was conducted at PBP in 2011 due to medical reasons. The end date for each field season is marked by an arrow in each panel. Here, precipitation inputs include snowmelt (*M*) and rainfall (*R*). Water losses comprise evaporation (including transpiration) (*E*), runoff (*Q*) while the residual is equivalent to ($\Delta S + \xi$) (see Equation (1)). The daily water balance for 2008 is not shown, as it is found in Young *et al.* (2010). Daily water budget information for Cape Bounty is not available.

intensity of these processes is especially obvious during the warm spring season of 2012. In cooler years, the residual term for LSC (Figure 9(a)) reflects snowmelt inputs as the basin does not respond readily to rainfall events and storage remains positive during 2009 and 2010. In the exceptionally warm, dry season of 2012, storage became negative early, as stream discharge drops off and higher evaporation rates in response to warmer air temperature occur. Later in the 2012 season, the residual term ($\Delta S + \xi$) begins to respond to rainfall (shows an upward trend) but evaporation losses stay high and so it remains negative.

Similar to LSC, snowmelt inputs and losses from discharge dominate the early season and help to drive down the residual term for WC (Figure 9(b)). The term remains negative in all seasons, but both in 2009 and 2012, it does start to respond to rainfall inputs in the later part of the season as evaporation rates drop off. It is not yet clear why the residual is much more negative in 2009 than in 2010 and 2012. One possibility is that the end-of-winter snow survey underestimated areas of deep snow in the basin (i.e., hillslopes with late-lying snowbeds, deep valley snow), and this unaccounted snow led to elevated discharge levels, which in turn caused the storage term to drop. Clearly, more effort is required to adequately assess the processes accounting for variations in the hydrology of these catchments.

Average water budgets (PBP, CBAWO and others)

Table 2 considers the average water budgets from all study sites in comparison to seasonal water budget investigations from several selected long-term runoff studies in arctic

Table 2 | Average seasonal water budgets, mm including Standard Deviation () for study sites and other small basins located in different Arctic and Subarctic environments

Basin	Description and area	Year	Sn or M	R	P = (Sn or M) + R	Q	E	ΔS + ξ	Q/P
High Arctic									
McMaster River, Cornwallis Island, 75°N 95°W; Woo (1983) quoted in Young & Woo (2004a)	Polar desert, barren ground and bedrock (33 km^2)	1976–1981	147(39)	43(18)	190(51)	158(29)	38(9)	−1(23)	0.84(1.29)
LSC, Bathurst Island, 75.7°N, 98.7°W; *This study*	Polar desert, barren ground, and tundra (0.2 km^2)	2008–2010, 2012	78(25)	77(19)	154(11)	53(46)	86(26)	+16(70)	0.34 (0.29)
WC, Bathurst Island, 75.7°N, 98.7°W; *This study*	Polar desert, barren ground and tundra (4.2 km^2)	2008–2010, 2012	66(30)	71(26)	137(10)	116(48)	83(23)	−62(42)	0.85(0.35)
East River Basin, CBAWO, Melville Island, 74.9°N, 109.5°W; *This study*	Polar desert, barren ground and tundra (11.6 km^2)	2007–2009	101(11)	53(15)	154(4)	63(8)	103(7)	−12(17)	0.41(0.06)
West River Basin, CBAWO, Melville Island, 74.9°N, 109.5°W; *This study*	Polar desert, barren ground and tundra (8.0 km^2)	2007–2009	82(3)	53(15)	135(12)	116(10)	103(7.0)	-84(29)	0.84(0.13)
Heather Creek, Ellesmere Island, 80°N, 84.5°W; quoted in Young &Woo (2004a)	Polar Oasis, tundra and bare ground (6.1 km^2)	1990–1991	72(37)	50(54)	122(17)	44(50)	68(47)	+4(23)	0.39(0.47)
Intensive Watershed, Devon Island, 76°N 85°W; Rydén (1977) quoted in Young & Woo (2004a)	Wetland, 63% vegetation and bare soil, 12% water, 25% bedrock (0.12 km^2)	1972–1974	137(29)	48(22)	185(18)	83(18)	81(25)	−15(30)	0.45(0.06)
Low Arctic									
Tiksi, Eastern Siberia, 71.7°N,128.8°E; Ishii *et al.* (2004)	Tundra, and rocky, lichen ground (5.5 km^2)	1997–1999	114(27)	135(71)	249(90)	223(102)	54(21)	−28(34)	0.87(0.09)
Havikpack Creek, NW Canada, 68.3°N, 133.5°W; Marsh *et al.* (2004)	Boreal forest (17.0 km^2)	1992–2000	148(23)	135(18	282(30)	110(15)	135(13)	+37(26)	0.39(0.04)
Imnavait Watershed, Alaska, 68.6°N, 149.4°W; (Kane *et al.* 2004)	Tundra (2.2 km^2)	1985–2003	120(33)	241(45)	359(51)	181(56)	178(30)	+1(32) only error term	0.50(0.12)
Subarctic									
Iittovuoma Basin, Northern Finland, 68.5°N, 21.4°E; Seuna & Linjama (2004)	Tundra, sporadic perma-frost, stony moraine, dwarf birch, low mountains (11.6 km^2)	1976–2002	N/A	N/A	574(100)	341(86)	230(79), Here, $E = (P - Q)$	N/A	0.60(0.11)
C3 sub-basin, Caribou-Poker Creek, Alaska, 65.2°N, 147.5°W; Bolton *et al.* (2004)	Boreal forest, 53% perma-frost (5.7 km^2)	1978–2003	131(70)	281(53)	412(80)	84(37)	202(20)	+137(66)	0.27(0.01) based on 3 years of data
Filiper Creek, Mogot, Eastern Siberia, 55.6°N 124.9°E; Vasilenko (2004)	Taiga-larch with birch and alder shrubs, moun-tains (4.7 km^2)	1976–1985	N/A	N/A	619	307	289	+38	0.50

Table modified after Young & Woo (2004a).
Sn = end of winter snowpack or snowmelt (*M*); *R* = rainfall; *P* = precipitation; *Q* = discharge; *E* = evaporation including transpiration; $\Delta S + \xi$ = the change in storage + error term (Equation (1)); and *Q/P* is the runoff ratio (discharge/total precipitation). *N/A* = not available.

regions of North America and Eurasia. These small stream basins include High Arctic to Subarctic sites and represent varying amounts of permafrost (continuous to sporadic), vegetation (polar desert to taiga, or none) and physiography (steep to low-lying terrain).

Snowmelt at PBP and CBAWO study catchments is lower than that measured in the McMaster Basin and on Devon Island (Intensive Watershed site), while summer rainfall receipt is much higher than that recorded at McMaster River during the mid-1970s to early 1980s. Higher rainfall amounts at PBP and CBAWO could possibly be an indication of intensification of the hydrologic cycle in this region, as a greater proportion of precipitation falls as rain instead of snow (Huntington 2006; Rawlins *et al.* 2010). Remarkably, snowmelt amounts at the study sites are now comparable to Heather Creek, a polar oasis basin located 25 km inland from Eureka, Ellesmere Island (Figure 1). This is an area which typically receives less snowfall than Resolute Bay. Overall, rainfall receipt at PBP and CBAWO is only slightly higher than that at Heather Creek.

Rawlins *et al.* (2010) indicate that snow depth has increased across Eurasia, while in North America the spring snow cover extent has decreased. In a recent modelling study, Bintanja & Selten (2014) suggest that precipitation will increase by 50% by the end of the 21st century in Arctic regions with most of it linked to local evaporation brought about by sea ice retreat. Despite much observational uncertainty due to the paucity of data in Arctic regions, future precipitation here is expected to peak in the fall and winter with most occurring as snow. At PBP and CBAWO, snow data results (Table 2) do not yet reflect this emerging trend. In relation to the other Arctic watersheds, rainfall still remains much lower for the QEI study sites owing to a cooler climate here relative to more southern latitudes (Kane & Yang 2004). For example, rainfall at Imnaviat Creek and Havikpack Creek, two Low Arctic basins, average 241 and 135 mm, respectively (Table 2).

Evaporation losses for the sites on Bathurst Island (LSC, WC) and Melville Island (East and West River basins) are much higher, generally double than early estimates for McMaster River basin in the mid-1970s. Interestingly, evaporation totals are either the same (LSC, WC) or slightly higher (East and West basins) than the intensive wetland site, Devon Island in the early 1970s. This pattern of enhanced water loss due to evaporation including transpiration reflects: (1) recent global warming across High Arctic regions (Woo & Young 2014) (see Figure 5); (2) a higher percentage of tundra vegetation in these study basins than the McMaster basin which enhances transpiration losses; and (3) a rising number of thaw days during the summer (June, July, August), i.e., the number of days with mean daily air temperature $>0\,^{\circ}$C. For instance, in the mid-1970s the number of thaw days at Resolute Bay weather station ranged from 50 to 80 (1976 to 1981), while in a recent 5-year span (2008 to 2012), a time period marked by the PBP and CBAWO studies, they reach 73 to 88. These latter years approach the high number of thaw days (86 to 92, 2008 to 2012) which are typical of the Eureka weather station (mean thaw days = 82, 1953–2012). This is a location often described as the *Garden Spot of the Arctic*. Despite warm/dry weather at Heather Creek, 25 km inland from Eureka, evaporation estimates are lower here than the study sites (Table 2). This reflects a short field season in the 1990s (usually May to early August), and the availability of surface moisture here. In the absence of rainfall, shallowly rooted tundra vegetation in polar oasis environments must shift strategies and depend on soil moisture storage supplied by ground ice melt (Edlund *et al.* 1990).

At the Eastern Siberia site Tiksi, mean evaporation is low and can be attributed to rocky, barren ground (Table 2). Liljedahl *et al.* (2011) recently remarked on the nonlinear controls on evapotranspiration in Arctic coastal wetlands and determined that there were multiple limitations to evapotranspiration which makes it difficult to formulate how wetland systems will respond to future warmer and wetter conditions. The same argument could be put forth for PBP and CBAWO and other high latitude watersheds, which are typically composed of heterogeneous ground: tundra, bare, rocky ground or low-lying wetlands in valleys.

In summary, we surmise, with some caution, that the recent longer thaw durations and higher losses of evaporation for our study sites in comparison to other past studies in the High Arctic are supportive of the hypothesis of intensification of the hydrologic budget for Pan-Arctic regions (Rawlins *et al*. 2010). However, there still remains much uncertainty in the measurement of evaporation (Huntington 2006) and the trends (e.g., Tiksi) of this water balance component and others across different Arctic regions and watersheds (Table 2).

Total runoff at the study sites is lower than McMaster River, Imnaviat Creek, Tiksi, Iittovuoma Basin, Filiper Creek but similar to others – Havikpack Creek, C3, Intensive Water shed Creek. Differences can be linked to a high winter snowpack (snowmelt) (McMaster River, Imnaviat Creek), a steeper gradient or well-defined rocky stream channel, which enhances runoff (Iittovuoma Basin, Filiper Creek, Tiksi). The average residual (change in storage, including error) is much more variable, differing between study sites and across the islands. At PBP, LSC shows an average positive storage; water is being held in the basin, while WC reveals a large negative storage, implying additional inputs of water are leaving the basin outlet. Deep valley snowpacks may have been initially underestimated during the snow survey in some years (Assini & Young 2012), or there was additional meltwater inflow from beyond the catchment boundary (Young & Woo 2004b). Ground ice melt contributions were not considered, so could definitely be another source of error.

This storage pattern differs at CBAWO, where both the East and West basins show negative values. Here, the West River basin's storage is more negative much like WC, and this pattern, like WC, could possibly be attributed to deep snow in channels and late-lying snowbeds; locations in the basin where snow is often underestimated during annual snow surveys (McLeod 2008, unpublished MSc thesis, Department of Geography, Queen's University, Kingston, Ontario). Akin to CBAWO and PBP stations, the residual term ($\Delta S + \zeta$) continues to show much variation across other Arctic watersheds too (Table 2). The Subarctic Boreal C3 sub-basin of Caribou-Poker Creek basin in Alaska has one of the highest estimates (+137 mm) tabulated, which implies that much water is held in the basin, likely soil moisture (Bolton *et al.* 2004), while at Imnavait Creek, Central Alaska, a site of tundra and continuous permafrost has the lowest error value (+1 mm, averaged over 19 years of record). Other stations report positive storages over the long term (Havikpack Creek: +37 mm; Filiper Creek: +38 mm), and others negative estimates (Intensive Watershed: −15 mm; Tiksi: −28 mm). In some sites (Northern Finland), storage or error is not even considered with evaporation calculated as the residual ($E = P - Q$) (Seuna & Linjama 2004). Some studies argue that improvement in this residual term will come from improved observation methods and calculation techniques for water balance components, especially soil moisture and evaporation (including transpiration) (Vasilenko 2004), while for other researchers, it might be better techniques to measure deep valley snowpacks, ground ice melt and water inflow from beyond basin perimeters (Young *et al.* 2010).

Table 2 indicates that runoff ratios are similar between WC, the West River and McMaster River catchment, while the ratios for LSC, East River, Imnaviat Creek and Trail Valley Creek are alike. The reasons for these patterns are not yet entirely clear and they may be linked to deep valley snow not being measured adequately, but the differences do deserve further attention. Average high runoff ratios exist for both WC (0.85) and West River (0.84) and are comparable to McMaster River (0.84) and Tiksi (0.87). Much lower runoff estimates for LSC (0.34) and East River (0.41) fall in the range (0.27 to 0.6) of other watersheds; sites situated from High Arctic to Subarctic locations (Table 2). Despite a similar climate, topography, vegetation, runoff ratios can differ widely between nearby basins (WC and LSC) or even adjacent ones (West and Easter Rivers). Clearly, understanding the landscape factors and processes controlling the streamflow regimes at these sites is critical if we are ever to estimate streamflow in other ungauged stations (Spence *et al.* 2005) across the QEIs, especially ones draining into the Northwest Passage.

CONCLUSIONS

1. The hillslope creeks at PBP, Bathurst Island and Cape Bounty, Melville Island are still exhibiting a nival-type regime although high discharge peaks are being generated by summer rainfall at Cape Bounty. Future expectations are that a nival regime will continue at these sites given that precipitation across the Arctic is expected to increase by 50% towards the end of the 21st century, with most of it falling as snow (Bintanja & Selten 2014).
2. Limited statistical analysis suggests an earlier onset in snowmelt runoff and peak snowmelt discharge at these catchments in comparison to previous streamflow studies

across the QEIs. However, caution should be exercised in the interpretation of these differences given the limited years of field data and study sites. Additional years of monitoring are clearly required and an expansion of the Water Survey of Canada hydrometric network in the QEIs is encouraged. Only then might it be possible to fully assess whether streamflow in this region is clearly responding to recent warmer spring/summer air temperatures with its apparent role in modifying the cryosphere through earlier snowmelt (Derksen & Brown 2012), dramatic loss of late-lying snowbeds (Woo & Young 2014), glacier mass reduction (Fisher et al. 2012; Zdanowicz et al. 2012; Lenaerts et al. 2013) and deeper active layers (Smith et al. 2010). Moreover, a comprehensive streamflow network across the QEIs would lead to improved understanding of future streamflow response to continued global warming. For instance, only with sustained and reliable field observations across a range of catchments will it be possible to determine whether an earlier and sustained start to runoff is actually occurring or prolonged baseflow, or that floods are becoming much more extreme than past conditions, all indications of intensification of the hydrologic cycle (Huntington 2006; Rawlins et al. 2010).

3. Finally, variations in streamflow and water budgets between catchments, across islands and with respect to other Arctic streamflow studies deserve further attention at this high latitude so as to better understand landscape factors and atmospheric processes modifying catchment hydrology, especially the storage + error term. Additional and detailed hydrograph analysis should prove insightful in this regard (e.g., pattern and duration of both the rising and recession limbs, magnitude of peak to low flows, shape of the hydrograph peak (sharp versus attenuated) and duration of flood/drought episodes, etc. (Carey & Woo 2000; Dingman 2000)).

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