Assessment of precipitation and snowcover in northern research basins*

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Abstract In 2004, a workshop was held to collect and synthesize the water balance data from 39 northern research basins (NRB) in Victoria, BC, Canada. One of the recommendations from the meeting was a need to review systematically each component of the water balance for these northern basins in order to identify spatial and temporal trends and to address significant knowledge gaps. Here, we assess the methodologies for measuring snow and rain in these northern basins; examine the temporal and spatial patterns of snow accumulation both during and at the end-of-the winter; consider ablation patterns and comment on the occurrence of extreme events. Our evaluation indicates that northern hydrologists still employ a variety of gauges and approaches to both measure and correct precipitation. For the NRB, rainfall contributions dominate in lower latitudes while snowfall gains importance with higher latitudes and altitude. Occurrence of large water bodies, topography (i.e. aspect, slope) and vegetation influence precipitation amount and its distribution across the landscape. Only two NRB studies showed a declining trend in snowcover (SWE). Snow is still considered the most important input of water in these northern basins, but extreme summer precipitation events (both rain and snow) have triggered higher magnitude floods than seasonal snowmelt runoff. Glacierized basins are sensitive to summer snowfalls and low winter snow storage. Both have the potential to dampen or enhance melting despite warmer or cooler air temperatures. Standardized gauges, approaches and continued monitoring of the NRB is encouraged.

Keywords Arctic environments; hydrology; northern circumpolar basins; rainfall; snowfall; water balance

Introduction

A recent Arctic Climate Impact Assessment report predicts significant changes for Arctic regions (ACIA 2005). Average arctic temperatures have already increased at almost twice the rate as the rest of the world and Arctic precipitation has increased by about 8% on average over the past century. Much of the increase has come as rain, with the largest increase in autumn and winter. It is expected that, by the end of the century, the Arctic as a whole will experience an annual total precipitation increase of roughly 20%, with most of this increase occurring as rain, accompanied with shorter and warmer winters (ACIA 2005). This future scenario remains unclear with increases in temperature not always resulting in higher precipitation (Przybylak 2002). Good estimates of precipitation are hampered by the sparseness of data in parts of the region (Serreze and Etringer 2003; ACIA 2005)

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and uncertainties in measuring it (mostly its solid form) due to undercatch errors (e.g. Aleksandrov *et al.* 2005). Furthermore, there has been a world-wide reduction in the number of stations throughout the circumpolar Arctic (New *et al.* 2001; Shiklomanov *et al.* 2002) and many of the existing weather stations are situated along coasts, at low elevations and close to communities and airports. Precipitation measured at these stations is known to underestimate snow amounts even in nearby basins (Woo *et al.* 1983a, b).

Snowfall is also difficult to measure accurately; with increasing wind speeds, snowfall catch is reduced. It has only been recently through the WMO precipitation gauge inter-comparison initiative that correction factors for both solid and liquid are starting to be applied systematically (Førland and Hanssen-Bauer 2000). These corrections have improved the reliability of precipitation records but problems still exist. For example, some researchers may not have access to wind and temperature records needed to correct precipitation data, or choose to just use raw data in their interpretations (e.g. Serreze and Etringer 2003). Blowing and drifting snow events still hamper measurements, with some records being discarded in the analysis. Precipitation usually occurs at the same time as blowing snow but it is often difficult to isolate the two components. There is also concern that observed increases in precipitation may in fact be the application of reduced correction equations for liquid precipitation (Førland and Hanssen-Bauer 2000). Trace events are those precipitation events which amount to less than or equal to 0.01 inches (ca. 0.2 mm) (National Weather Service 1998), and contribution to annual totals are still being ignored in some regions. In northern Norway, trace events amounted to 20% of the total events, but such events were still only given a value of 0 mm (Førland and Hanssen-Bauer 2000).

Some researchers overcome the difficulties of snowfall measurements by measuring the snowpack water equivalence (SWE) at the end-of-the winter period, since midwinter melt events are considered rare in this region and meltwater is generally the dominant mode of water into Arctic basins. In some cases, snow indices based on terrain unit or elevation range are developed in relation to nearby weather station data to provide an approach to extend point data over space (Woo and Young 2004). This initiative improves snowcover estimates for a basin but is time-consuming, often demanding yearly fieldwork. While remote sensing imagery can adequately define the extent of snowcover and its distribution, its application to northern regions is still hampered by inadequate algorithms to define SWE, long return times of satellites (e.g. 16 d for LANDSAT), and lack of validation. Remote sensing imagery also does not extend back beyond the 1970s (New *et al.* 2001), requiring researchers to analyze weather station time series data and interpolate this information over large geographical areas.

In March 2004, a Northern Research Basins (NRB) Workshop was held in Victoria, BC, Canada. A northern research basin is defined here as a research basin located in or near the circumpolar north, in which snow (or snowmelt runoff) is typically the dominate hydrologic process. The aim of the meeting was to synthesize water balance information for a range of experimental basins existing throughout the circumpolar north and to examine in more detail individual components of the water balance framework. The purpose of this paper addresses the latter objective and draws heavily from water balance information contained in the *IASH Publication* 290 (Kane and Yang 2004a) which arose from the workshop findings. Where possible, we supplement this information with recent precipitation studies of northern regions. Therefore, the objectives of this paper are to: (1) assess the methodologies for measuring snow and rain in these northern basins; (2) examine and discuss the pattern of snow accumulation and rain on a regional and temporal basis; (3) consider ablation patterns and (4) comment on the occurrence of extreme events. The report attempts to clarify the spatial and temporal variations of snow and rain existing in monitored northern basins, and elucidate the factors accounting for observed differences. Finally, we document the shortcomings of our present precipitation research efforts and indicate the direction of future hydrological focus.

Methodologies for measuring snow and rainfall

Instrumentation to measure snow and rainfall in the NRB is not yet standardized. Employment of precipitation equipment (both recording/non-recording; shielded/non-shielded gauges) often depends on familiarity with the instrumentation, the focus of the study, the ease of accessibility to and within the study basin and availability of meteorological information to correct precipitation records, specifically wind and temperature data. Russian basins (e.g. Mogot–Vasilenko 2004; Nizhndevidk – Zhurvin 2004b; and Valdai–Shutov 2004) are typically well instrumented (manual gauges, Tretyakov gauges (shielded and unshielded)), the gauges are visited frequently (daily, weekly to monthly) and measurements are corrected (wind, wetting, evaporation). Finnish basins have used a variety of gauges over the years. Until 1981, a Wild raingauge was employed, from 1981–1992, a Tretyakov gauge and from 1992 till now a Holmberg and Heiskanen gauge, a modification of the Tretyakov gauge. Finnish researchers, Seuna and Linjama (2004) have been diligent in the correction of their long-term data set in terms of correcting for site and gauge changes, in addition to variable atmospheric conditions. Application of the bias correction methods derived from the WMO gauge intercomparison (Goodison *et al.* 1998) can result in significantly higher estimates of precipitation, particularly in high latitude regions. For instance, these corrections have increased the winter and annual precipitation amounts by up to 50–100% in Siberia (Yang and Ohata 2001) (Figure 1). Førland and Hanssen-Bauer (2003) found that for six Arctic Norwegian stations, it can vary from 148 to 166% of the observed precipitation on an annual basis. Lowest corrections are found for Island stations (Jan Mayen and Bjørnøya) and the highest for more inland sites on Svalbard. For basins which are not visited during the winter months, precipitation gauges are not routinely employed (e.g. Canadian High Arctic (Young and Woo 2004a, b); Canadian Low Arctic (Marsh *et al.* 2004); and Alaskan Low Arctic (Bolton *et al.* 2004)). Instead the researchers carry out extensive stratified snow surveys prior to snowmelt. Because of extremely windy conditions in glacierized basins, most glaciologists use unscreened gauges (e.g. Pisissurfik glacier, Greenland (Helweg 2004)) and/or use snow depth sensors to help correct coastal weather station data (e.g. Mittivakkat glacier, Greenland (Hasholt and Mernild 2004)). But since corrections may only apply to a certain windspeed range ($\leq$ 8 m/s-snow; $\leq$ 15 m/s-rain) large precipitation errors still occur at the Mittivakkat glacier, especially during blizzard conditions.

Areal distribution of precipitation (extrapolation from point to area)

Precipitation measurements are usually point measurements, measured in gauges with a typical orifice area of 200–400 cm^2. Hydrologists, however, need to know average or areal precipitation on a catchment scale, where the catchment area may be from a few km^2 up to

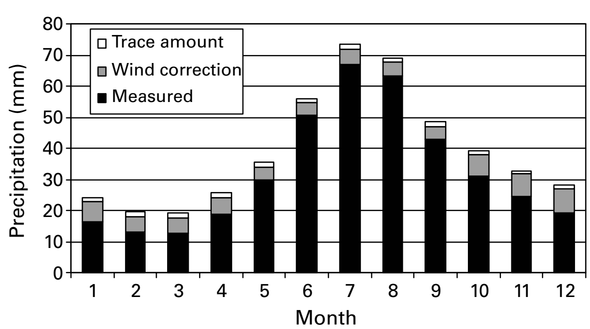

Figure 1 Overall mean precipitation (mm, measured, corrected, trace) for 61 climate stations in Siberia, 1986–92 (source: Yang and Ohata 2001)

several thousand km^2, always many orders of magnitude larger than the gauge area. The general problem is therefore to estimate the spatially averaged areal precipitation from one or more point precipitation measurements. Many different methods have been developed both for computing areal precipitation and the spatial precipitation distribution, each with different strengths and weaknesses. Weighted average approaches assign a weighting to each precipitation gauge and include the arithmetic-average method, the Thiessen polygon method and the Bethlamy two-axis technique (Dingman 1993). In surface-fitting methods the measured values are used to identify a continuous surface which is used to represent the precipitation $P(x,y)$ at all points of interest in the catchment. This surface is then used to describe the spatial distribution of precipitation, for example, for a single storm, a daily or annual sum. Surface-fitting methods vary from the classical isohyetal method where fitting is done by eye to complex methods like kriging or empirical orthogonal functions (EOF) with high computational complexity (e.g. Dingman 1993). Among other methods the hypsometric method is commonly used, but the method is only useful for catchments with orographic precipitation, where the precipitation is a function of elevation. Appropriate lapse rates need to be determined since regional variations exist (Killingtveit *et al.* 2003). For example, for Norwegian glacierized basins (e.g. Bayelva and De Geerdalen catchments), elevational corrections range from 15–20% per 100 m respectively (Killingtveit 2004). In Sweden precipitation gradients of 5–10% per 100 m are recommended for use in design flood computations (Bergström *et al.* 1992) and Linacre and Geerts (1998) state that precipitation at Resolution Island, just south of Baffin Island, increases 9% per 100 m, while at White Glacier on Axel Heiberg Island (79°N), it is only 7% per 100 m.

All the different methods rely on a reasonable number of gauging stations inside or close to the catchment, in order to compute the areal precipitation with reasonable accuracy. This is a problem in many Arctic catchments where the number of precipitation stations in operation is usually very low. The recommended minimum density of precipitation stations in polar regions is only 1 station per 10,000 km^2 (WMO 1994). This may be compared to the recommendations for plains and hilly/undulating areas (575 km^2) and mountainous catchments (250 km^2).

In summary, most northern research basins conduct routine snow surveys across diverse landscapes in their basins to provide either a check on point snowfall measurements (e.g. Seuna and Linjama 2004; Zhuravin 2004a, b) or to arrive at a realistic estimate of snow amount for a remote basin where precipitation gauges are visited infrequently and/or coastal weather stations are not considered truly representative of the surrounding terrain (e.g. Young and Woo 2004a, b and others.). The summer precipitation raingauge network is more detailed, visited more frequently though consideration and treatment of errors is not always clearly documented.

Spatial and temporal patterns of snow and rain

Precipitation (snow and rain) can generally be regarded as the most important input of water into a basin. Given that precipitation is one of the most difficult components of a water budget to be estimated adequately and record length can be episodic for many northern experimental basins, only general patterns will be stressed here.

The circumpolar north is a diverse area comprising a variety of physiographic regions. The water balance studies reviewed (see Kane and Yang 2004b) range in climate from Temperate, Subarctic, Low Arctic to High Arctic; in ecosystem types (boreal forest, wetland, taiga, tundra, and polar desert); in elevation (sea level to 2000 m); and in permafrost distribution (permafrost free to continuous). Some basins are highly glaciated, while others have various amounts of glacial ice or lingering snowdrifts which tend to supplement stream discharge during the warmer months (see Kane and Yang 2004a, b). Clearly, given the high variability in environments across the circumpolar north, it is not surprising that large spatial and temporal variations in snow and rain are observed.

Spatial

Kane and Yang (2004b) list 39 study basins, their locations, period of study and water balance components. This information together with the published water balance papers in *IAHS Publication* 290 is utilized in Figures 2 and 3 to assess general patterns in precipitation and rainfall. Here, for discussion purposes we loosely define High Arctic basins as > 70°N, Low Arctic-Continental and -Maritime-type basins range from 60–70°N and Subarctic basins extend from 50–60°N. Glacerized basins are coupled together, and there is one temperate mountainous basin (ca. 44°N). To facilitate the discussion, information about each basin's location, area and length of record considered in this report is included in Table 1. This information follows from Kane and Yang (2004b).

Average annual precipitation

Average annual precipitation is plotted in Figure 2. High Arctic basins, in general, have the lowest annual precipitation amounts (ca. 200 mm/yr) followed by Low Arctic and Subarctic basins (ca. 400 and 600 mm, respectively). The greatest annual precipitation amounts arise in glacierized basins (> 600 mm) (e.g. Hasholt and Mernild 2004; Killingtveit 2004) and in the one mountainous temperate basin (Moshiri watershed) located in Northern Japan (1600 mm) (Ishii *et al.* 2004b) owing to the orographic effect on precipitation (Killingtveit 2004). Overall, this pattern is reasonable, given the drier and colder environments of High Arctic environments (Young and Woo 2004a, b). Generally, for glacier basins, snow increases with altitude, reflecting the longer accumulation period and increased precipitation at higher elevations (Winther *et al.* 1998). Accumulation is also modified by topography with low values measured on more level areas above steep slopes and high accumulation below steeper sections in a down-glacier direction (Hasholt and Mernild 2004). Both Low and Subarctic basins are warmer and moister. The Finnish basins (Low Arctic-Maritime) have

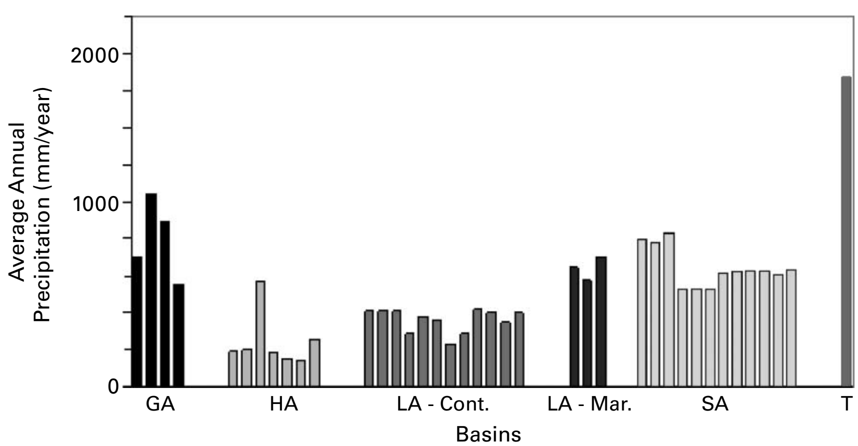

Figure 2 Average annual precipitation (mm/year) for Northern Research Basins considered in Kane and Yang (2004*a*). Here **G** refers to glacierized basins (left to right: Pisissarfik, Mittivakkat, Bayelva, Svalbard, De Geerdalen, Svalbard); **HA** indicates High Arctic basins (left to right: McMaster River, Gully River, Ross Point, Intensive Watershed, Hot Weather Creek, Heather Creek, Tiksi); **LA-Cont**. refers to Low Arctic basins having continental-type climates (left to right: C2 (CPCRW), C3 (CPCRW), C4 (CPCRW), Wolf Creek, Upper Kuparuk (Kuparuk), Imnavait Creek (Kuparuk), Trail Valley Creek (Mackenzie), Havikpak Creek (Mackenzie), Scotty Creek, Kantakovy Creek (Kolyma), Yuzhny Creek (Kolyma), Severny Creek (Kolyma)); **LA –Mar**. refers to Low Arctic basins having maritime influences (left to right: Vähä-Askajoki, Iittovuoma, Laanioja); **SA** refers to Subarctic basins (left to right: Log Usadievsky, Log Taezhny, Polomet River, Dead Creek, Teako Creek (Dead Creek), Wild Goose Creek (Dead Creek), Nelka River (Mogot), Zakharenok River (Mogot), Filiper River (Mogot), Yasenok (Nizhnedevitsk), Devista River (Nizhnedevitsk), Dolgy Ravine (Nizhnedevitsk) and **T** is the one temperate basin (Moshiri). See Table 1 for more details on the NRB

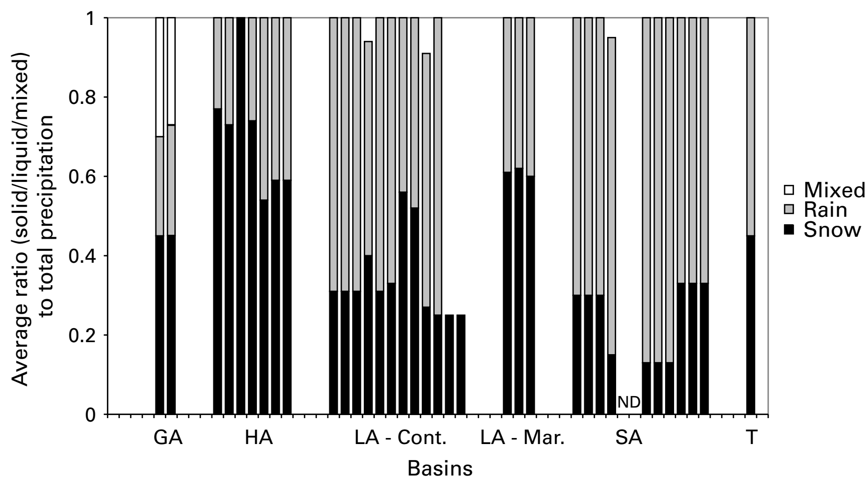

Figure 3 Average ratios of solid (snow: solid black); liquid (rain: grey) and mixed (open) to total precipitation for the Northern Research Basins considered in Kane and Yang (2004a). Here **G** refers to glacierized basins (left to right: Bayelva, Svalbard, De Geerdalen, Svalbard). **HA** indicates High Arctic basins (left to right: McMaster River, Gully River, Ross Point, Intensive Watershed, Hot Weather Creek, Heather Creek, Tiksi); **LA-Cont**. refers to Low Arctic basins having continental-type climates (left to right: C2 (CPCRW), C3 (CPCRW), C4 (CPCRW), Wolf Creek, Upper Kuparuk (Kuparuk), Imnavait Creek (Kuparuk), Trail Valley Creek (Mackenzie), Havikpak Creek (Mackenzie), Scotty Creek, Kantakovy Creek (Kolyma), Yuzhny Creek (Kolyma), Severny Creek (Kolyma)); **LA – Mar**. refers to Low Arctic basins having maritime influences (left to right: Vähä-Askajoki, littovuoma, Laanioja); **SA** refers to Subarctic basins (left to right: Log Usadievsky, Log Taezhny, Polomet River, Dead Creek, Teako Creek (Dead Creek), Wild Goose Creek (Dead Creek), Nelka River (Mogot), Zakharenok River (Mogot), Filiper River (Mogot), Yasenok (Nizhnedevitsk), Devista River (Nizhnedevitsk), Dolgy Ravine (Nizhnedevitsk) and **T** is the one temperate basin (Moshiri). Note values of 0 within categories indicate insufficient data provided to assess ratios for basins. Owing to missing information about snow totals, rain or mixed events, some ratios do not sum to 1

slightly higher precipitation amounts ($>$ 600 mm/yr) than their Continental counterparts owing to their close proximity to the Barents Sea (Kane and Yang 2004b). Within each category, variations do occur and likely can be attributed to the local physiography, climate, and nearness to a water source and basin size. For example, one small catchment (0.1 km^2) in the Canadian High Arctic receives about 600 mm/yr. This particular catchment near Ross Point, Melville Island is fed by a large late-lying snowbed, which provides meltwaters throughout the summer period (Young and Woo 2004a). The Valdai wetland catchments, located in a Russian subarctic environment, receive about 200 mm more than other subarctic sites. These basins are considered to have continental climates but tend to be wetter than other sites (Kane and Yang 2004b).

Ratio of solid precipitation (snow) to total precipitation

Figure 3 shows the proportion of snow (solid) to total precipitation for most of the basins. It is evident in the High Arctic environments most of the precipitation input is in the form of snow. For most sites in this zone, snow accounts for $>70\%$ of the total precipitation, with less at Hot Weather Creek and Heather Creek, located in the Eureka intermontaine region, Ellesmere Island, Canada. Both of these latter sites experience a precipitation-shadow effect as cyclones tracking from the Arctic Ocean are deflected by the mountains along Axel Heiberg Island (Young and Woo 2004a). There is not a large difference in snow amounts between Low Arctic and Subarctic environments, though variability between sites does occur. Snow comprises about 30% of the total precipitation for these regions. Higher snow totals $>50\%$ occur at Trail Valley Creek and Havikpak Creek, near Inuvik, NWT, Canada,

Table 1 Research basins examined in this study along with available information for location, size, period of study and researcher (s). (Source: Kane and Yang 2004b)

Basin	Country	Latitude	Longitude	Area (km^2)	Period of study	Researcher(s)
Glacier						
Pisissarfik	Greenland	67.1°N	52.8°W	32.5	1996–2000	Helweg 2004
Mittivakkat	Greenland	65.7°N	37.8°W	20.0	1998–2002	Hasholt and Mernild 2004
Bayelva, Svalbard	Norway	78.9°N	11.9°E	30.9	1990–2001	Killingtveit 2004
De Geerdalen, Svalbard	Norway	78.3°N	11.3°E	79.1	1990–2001	Killingtveit 2004
High Arctic						
McMaster River	Canada	75.0°N	95.0°W	33	1976–1981	Young and Woo 2004b
Gully River	Canada	76.0°N	85.0°W	22.4	1972–1973	Young and Woo 2004b
Ross Point	Canada	75.0°N	107.3°W	0.1	1986	Young and Woo 2004b
Intensive Watershed	Canada	76.0°N	85.0°W	0.1	1972–1974	Young and Woo 2004b
Hot Weather Creek	Canada	80.0°N	84.5°W	130.0	1989–1991	Young and Woo 2004b
Heather Creek	Canada	80.0°N	84.5°W	6.1	1990–1991	Young and Woo 2004b
Tiksi	Russia	71.7°N	128.8°E	5.5	1997–1999	Ishii *et al.* 2004a
Low Arctic-Continental type climate						
C2 (Caribou-Poker Creek Watershed-CPCRW)	USA	65.2°N	147.5°W	5.2	1978–2003	Bolton *et al.* 2004
C3 (CPCRW)	USA	65.2°N	147.5°W	5.7	1978–2003	Bolton *et al.* 2004
C4 (CPCRW)	USA	65.2°N	147.5°W	11.4	1980–2003	Bolton *et al.* 2004
Wolf Creek	Canada	61.0°N	135.0°W	195.0	1993–1994, 1995–1996	Janowicz *et al.* 2004
Upper Kuparuk (Kuparuk)	USA	68.6°N	149.4°W	142.0	1996–2002	Kane *et al.* 2004
Imnavait Creek (Kuparuk)	USA	68.6°N	149.4°W	2.2.0	1995–2002	Kane *et al.* 2004
Trail Valley Creek (Mackenzie)	Canada	68.7°N	133.5°W	68.0	1991–2000	Marsh *et al.* 2004
Havikpak Creek (Mackenzie)	Canada	68.3°N	133.5°W	17.0	1991–2000	Marsh *et al.* 2004
cotty Creek	Canada	61.3°N	121.3°W	152.0	1999–2002	Quinton *et al.* 2004
Kantakovy Creek (Kolyma)	Russia	61.9°N	147.4°E	21.2	1970–1985	Zhuravin 2004a
Yuzhny Creek (Kolyma)	Russia	61.9°N	147.4°E	0.3	1970–1985	Zhuravin 2004a
Severny Creek (Kolyma)	Russia	61.9°N	147.4°E	0.4	1970–1985	Zhuravin 2004a
Low Arctic-Maritime influenced climates						
Vähä-Askajoki, 114	Finland	66.6°N	27.6°E	16.4	1958–2003	Seuna and Linjama 2004
Iittovuoma, 117	Finland	68.8°N	25.4°E	11.6	1977–2003	Seuna and Linjama 2004

Table 1 – *continued*

Basin	Country	Latitude	Longitude	Area (km^2)	Period of study	Researcher(s)
Laanioja, 121	Finland	66.6°N	27.6°E	16.4	1976–2003	Seuna and Linjama 2004
Subarctic						
Log Usadievsky	Russia	57.6°N	33.1°E	0.4	1967–1985	Balonishnikova *et al.* 2004
Log Taezhny	Russia	57.6°N	33.1°E	0.5	1967–1985	Balonishnikova *et al.* 2004
Polomet River	Russia	57.6°N	33.1°E	432.0	1967–1985	Balonishnikova *et al.* 2004
Dead Creek	Canada	50.0°N	95.0°W	106.0	1982–2005	Thorne and Hawkins 2004
Teako Creek (Dead Creek)	Canada	50.0°N	95.0°W	0.5	1982–2005	Thorne and Hawkins 2004
Wild Goose Creek (Dead Creek)	Canada	50.0°N	95.0°W	0.4	1983–2005	Thorne and Hawkins 2004
Nelka River (Mogot)	Russia	55.6°N	124.9°E	30.8	1976–1985	Vasilenko 2004
Zakharenok River (Mogot)	Russia	55.6°N	124.9°E	5.8	1976–1985	Vasilenko 2004
Filiper River (Mogot)	Russia	55.6°N	124.9°E	4.7	1976–1985	Vasilenko 2004
Yasenok (Nizhnedevitsk)	Russia	51.5°N	38.4°E	21.7	1973–1988	Zhuravin 2004b
Devista River (Nizhnedevitsk)	Russia	51.5°N	38.4°E	103.0	1973–1988	Zhuravin 2004b
Dolgy Ravine (Nizhnedevitsk)	Russia	51.5°N	38.4°E	2.6	1973–1988	Zhuravin 2004b
Temperate						
Moshiri	Japan	44.4°N	142.3°E	1.2	1988–1998	Ishii *et al.* 2004b

likely owing to their proximity to the Beaufort Sea and cold prevailing temperatures (i.e. high northern location ~68°N) (Marsh *et al.* 2004). The importance of a nearby water source is also evident for the Finnish basins (Low Arctic-Maritime). These basins lie at 68–69°N, but their close proximity to the Barents Sea results in snow accounting for 60% of the annual precipitation total, about 30% greater than most other Low Arctic sites (Seuna and Linjama 2004). The Moshiri watershed in Northern Japan (lat. 44°N, elevation range: 280–545 m a.s.l) experiences orographic precipitation with an unlimited moisture supply coming from the Pacific Ocean. Snow accounts for about 45% of the total precipitation input (Ishii *et al.* 2004b).

This spatial pattern of solid precipitation is supported by other studies. In a recent investigation of long-term precipitation and snow depth from weather stations across the Barents and Kara seas, Aleksandrov *et al.* (2005) indicate that for the northern and northeastern parts of this region, 50–70% of the annual precipitation is solid; while in the southern areas the solid fraction is only 35–43%.

Figure 3 also provides an indication of the proportion of mixed, liquid to total precipitation for most of the non-glacerized basins. In the Low Arctic and Subarctic environments, rain accounts for about 70% of the annual precipitation, although variations occur within each region. For Low Arctic sites, the greatest fraction of rain occurs for the Kantakovy Creek basin, Kolyma, Russia (~76%, lat. 62°N) (Zhuravin 2004a), while the smallest occurs for Trail Valley Creek and Havipack Creek near Inuvik, NWT, Canada (44 and 48%, respectively) (Marsh *et al.* 2004). This is expected since the latter sites lie further north (68°N). Rain is not as important for the Finnish basins (Low Arctic-Maritime) as is snow, comprising only about 40% of the total (Seuna and Linjama 2004). For Subarctic sites, the greatest proportion of rain occurs for the Mogot basins in Russia, which are situated at about 55°N (Vasilenko 2004). Greatest variability in rainfall occurs in High Arctic environments, where its importance can vary from about 23% (McMaster River Basin, Cornwallis Island, Canada) (Young and Woo 2004a) to 55% (Tiksi, Siberia) (Ishii *et al.* 2004a). Insufficient information exists for mixed events at most study sites and so these events are not reported here.

Temporal

Temporal information relating to rainfall and snowfall occurrence is sparse, but some general patterns do emerge. For High Arctic sites, snow tends to occur in September and lies on the ground until late May or June, at which time the main snowmelt period occurs. For Low Arctic sites, there is some variation in the timing and duration of the snowcover. For selected basins, snow stays on the ground from September to March or April, while for other basins, snow does not appear until about October but lasts until May (e.g. Kolyma basins, Russia). Understandably, the Subarctic basins have the shortest duration of snowcover. Here snow generally does not occur until October or November and most disappears by March or April. At these lower latitudes mid-winter thaws can influence the final snowpack SWE (Zhuravin 2004b). For instance, at the Nizhnedevitsk basin the maximum snow water equivalent during the period under consideration varied from 34 to 124 mm with the snowcover accumulation regime being influenced significantly by numerous winter thaws. The Moshiri watershed also has a short snowcover period with snow occurring in October and disappearing in March.

In the context of climate change, defining trends in either increases or decreases in snow amounts across circumpolar regions is of great interest and importance. For all of the NRB, the precipitation and SWE record is short (e.g. Bolton *et al.* 2004; Ishii *et al.* 2004a, b; Marsh *et al.* 2004; Thorne and Hawkins 2004; Young and Woo 2004a) and no credible trends are apparent with much year-to-year variability arising. However, basins with longer term records do show some signs of change. Seuna and Linjama (2004) have found that there is a slight decrease in precipitation for the southern basin of Hovi, Finland (62°N). During the period 1976–2002 the basin shows a general decline in maximum SWE. This pattern differs from the northern Finnish basin of Vähä-Askanjoki (66°N) where no trend in the SWE record is yet apparent, a finding which is supported by Aleksandrov *et al.* (2005). They found no declining trend in maximum snow depth for southwestern weather stations of the Barents–Kara seas region, the same sector where the Vähä–Askanjoki basin would lie. However, they did find maximum snow depths (April or May) declining at northeastern stations at a rate of 8–9 cm per 40 years (1951–1992). Long-term records in Russia probably provide the best indication for change. Figure 4 indicates a clear descending trend in SWE at Valdai (in the Usadyevsky catchment) which is comparable to other locales in European Russia (Shutov 2000).

While snowfall can occur in any month of the year, most rainfall occurs during the warm period after snowmelt and prior to freeze-back. All NRB (Kane and Yang 2004a) demonstrate considerable variability in rainfall both on a seasonal and annual basis. Recorded maximum average rainfalls for High Arctic sites range from 67 mm (McMaster Basin, Canada) (Young and Woo 2004a) to 217 mm (Tiksi, Siberia) (Ishii *et al.* 2004a). For Low Arctic basins, it ranges from 165 mm (Havikpak Creek, Canada) (Marsh *et al.* 2004) to 495 mm (Kantakovy Creek, Kolyma, Russia) (Zhuravin 2004a).

Snow ablation pattern

All northern basins share a number of similarities in terms of snow ablation. Once the snowpack warms up and becomes isothermal then melt and loss of the snowcover occurs quite rapidly. For the most part, the snow ablation period except for mountainous and glaciated basins (e.g. Sharp *et al.* 2002; Helweg 2004) is generally over in a two or three week period. All sites share a distinctive daily rhythm of melt, with melt intensity being largely driven by net radiation. As the snow becomes patchy, advective sensible heat fluxes tend to become more important. For all sites and within sites, the timing of this event, its duration and daily melt rates can vary from year to year.

Figure 5 from Alaska provides a good illustration of annual snow ablation. Variations do exist including the onset and duration of the melt period and this is typical for all regions. Snowmelt often starts earlier in the intermontaine region of Ellesmere Island (80°N) than it does for more southern latitudes. Mountains on Axel Heiberg shelter the area from adverse

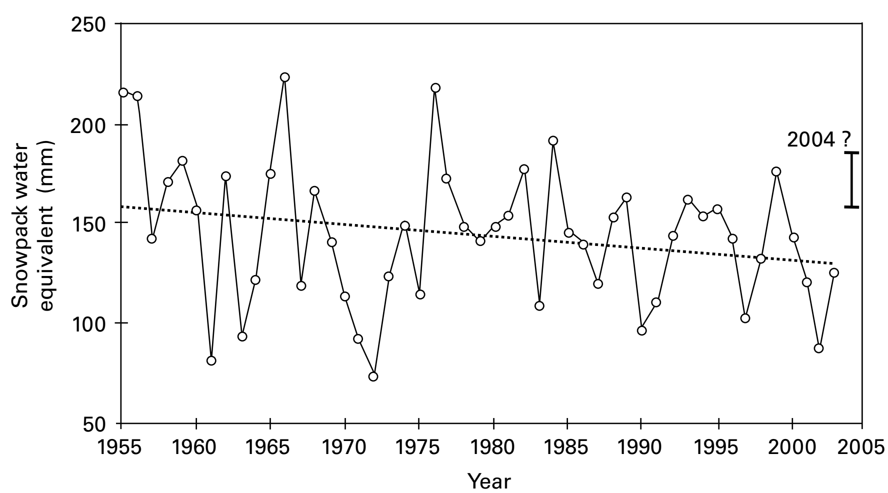

Figure 4 Maximum snow water equivalents (SWE, mm) prior to snowmelt at Valdai, Russia. The data here were obtained on a special runoff plot with an impermeable (concrete) surface. "?" indicates that 2004 data have been estimated

weather conditions and this sheltering effect enables early warming so that the beginning of snowmelt is often advanced by about a month ahead of the polar desert areas further south (Woo and Young 1997). For small catchments, the presence of large and persistent snowdrifts may supply meltwater long after the seasonal snowcover is depleted. Lewkowicz and Young (1991) found that the ablation of a large late-lying snowbed near Ross Point, Melville Island yielded the bulk of water (535 mm) to a small 0.1 km^2 basin. At the local scale, there are differences in the response of snowmelt from slopes which can affect runoff patterns and timing of streamflow. Janowicz *et al.* (2004) indicates snow ablation varied between north- and south-facing slopes in the Wolf Creek Basin (61°N). Snow ablation on the north-facing slopes proceeded slowly for the span of 5 weeks, yet produced significant runoff due to the relatively impervious permafrost and limited storage in the active layer substrate. Snowmelt on the south-facing slope, containing no permafrost, was rapid and complete one month prior to the north-facing slope. No runoff was observed with a subsequent large increase in storage change. With respect to glacierized basins, Knuden and Hasholt (2003) report that for the Mittivakkat glacier, located in southeastern Greenland, ablation generally decreases with height. This reflects the longer ablation period at low altitudes and the diminishing net energy supply with height. They observe that ablation decreases strongly with altitude up to about 500 m (the level of equilibrium line altitude, ELA). At this level, the duration of snow cover increases along with cloud cover which helps to dampen radiation receipt.

Extreme events

Extreme events can take many forms: rain or snow, mid-winter melt, heavier or lighter than normal snowpacks and droughts (Kane *et al.* 2003a). It is difficult to comment on the occurrence of extreme events for NRB basins since, in most cases, precipitation records are short. Yet all researchers agree that while these events are variable both spatially and

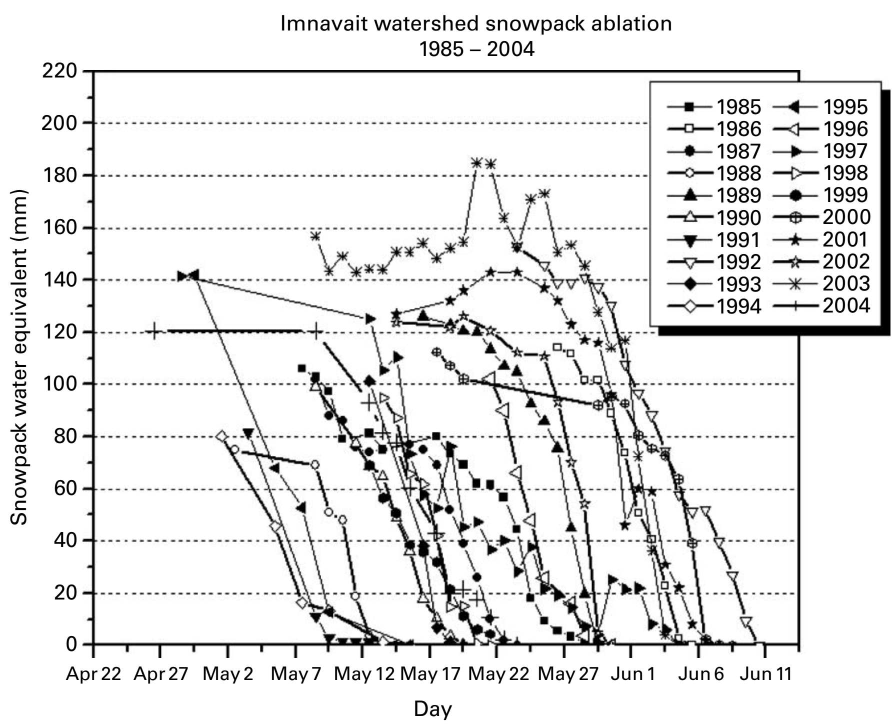

Figure 5 Snow ablation pattern for the Imnavait watershed, Arctic Alaska from 1985–2004

temporally, they are hydrologically significant and serve to transport large volumes of sediment and modify drainage networks (Kane *et al.* 2003a, b). For the most part, the end-of-winter snowpack generates the largest flood for the season. But lately, northern researchers (e.g. Kane *et al.* 2003a, b) have been reporting the occurrence of large summer precipitation events (both snow and rainfall) which are triggering high runoff and stream discharges much larger than the annual snowmelt regime. Kane *et al.* (2003a) indicate that, for one event in Arctic Alaska, a 100 m^3/s peak flood was generated 3 × higher than the previously recorded snowmelt flood. The magnitude of these events can be made more severe because of topography (steep ground); landscape (rock vs. % of lakes/ponds) along with frost table position and presence of permafrost which limits the subsurface storage capacity.

Kane *et al.* (2003a) also indicate that a few intense precipitation events appear to generate greater runoff than numerous less intense rain storms. However, it is the minor rainfall events which may be important in priming a basin for the high magnitude events by first satisfying water storages.

For glaciated and ice-covered basins, low snow years may lead to extreme melting. A good example is provided by the recent ablation event on the southwest slope of the Greenland ice sheet in 2003. The period from September 2002 to April 2003 was characterized by a high pressure system over the Norwegian Sea and a low pressure system on the southeast tip of Greenland. This pattern resulted in large and frequent storms to southeastern Greenland. Fewer storms in west Greenland and precipitation shadow effects caused little precipitation to fall along the southwest slope. Subsequently, a shorter snowmelt period and a longer ice ablation period (lower albedo) resulted for this area. Record high melt rates occurred and overall 71 km^3 more melting and freshwater discharge than normal occurred from the Greenland ice sheet. The majority of this discharge originated from the southwest sector (Box *et al.* 2005).

This critical control of snow (either winter or summer precipitation) is also reported for the Mittivakkat glacier, southeastern Greenland (see Knuden and Hasholt 2003). Despite 1998 having the lowest mean temperature, it had the highest amount of ablation on record. This was attributed to the low summer precipitation–snowfall coverage over the glacier surface. Conversely, 1996 had the highest mean air temperature but lowest ablation. This time, a record high winter snow balance was able to maintain a high albedo throughout the summer period, thereby minimizing melt. Clearly, the percentage of snow versus ice seems to be the major determinant of meltwater generation in these northern glacierized basins.

Summary and conclusions

Precipitation (snow and rain) is an important component of the terrestrial water balance controlling moisture storage and release. It is also an important link to the climate system through its influence on surface reflectance and by its control of heat and moisture transfers to the atmosphere from the underlying ground surface (Brown 2001). This assessment of the precipitation term (P) as measured in NRB stations, while qualitative in scope, does permit a number of generalizations to be made. First, precipitation amounts are generally higher at low latitudes than high latitudes and the importance of snow increases with latitude and elevation. The occurrence of a nearby water source (e.g. Finnish basins – Barents Sea (Seuna and Linjama 2004), Moshiri – Pacific Ocean (Ishii *et al.* 2004b) and/or the local control of terrain (e.g. Janowicz *et al.* 2004; Young and Woo 2004a, b) and vegetation (e.g. Shutov 2004) further modify rain and snowfall amounts. All basins show a high degree of year-to-year variability and, except for one or two basins with long-term records, no explicit long-term trend in precipitation (rain/snow) can be detected. Snow accumulation at most basins begins in the fall and can last well into June and early July. For low latitude basins, winter thaws can lower winter SWE values but for most basins, because of persistent cold temperatures, spring thaws are uncommon and maximum SWE occurs just prior to the spring snowmelt period (March to mid-June). Once snowpacks are ripened, snow ablation occurs quickly and is generally finished in two or three weeks. Snow ablation lingers in both glacerized basins and in mountainous terrain possessing ice fields and/or late-lying snowpatches.

At this time, it is difficult to assess precipitation errors since studies use a variety of techniques and methodology. While some studies employ lapse rate relationships and/or corrected point gauge precipitation data and then spread it spatially over their basins (e.g. Killingtveit *et al.* 2003; Killingtveit 2004) other studies do not. Many studies avoid the inherent problems of point precipitation data (i.e. sublimation, blowing snow) instead by obtaining end-of the year winter snow (SWE) for their basins (e.g. Kane and Yang 2004b; Marsh *et al.* 2004; Young and Woo 2004a, b). However, techniques and approaches in obtaining end-of-year SWE values can also differ amongst scientists as can ablation measurements. Bolton *et al.* (2004) rely on direct ablation measurements to derive daily SWE values for a range of terrain in their basins, Arctic Alaska, while others utilize physically based surface energy balance models to deplete the winter snowcover (Woo and Young 2004).

What is clear from examining basin studies and recent modeling efforts is that reliable and consistent precipitation records for mountainous and steep terrain have still not been achieved (e.g. Bowling *et al.* 2003; Woo and Young 2004). This may be attributed to the insufficient number of precipitation gauges, logistical difficulties of getting to these sites and/or our incomplete understanding regarding both sublimation and blowing snow processes (Liston and Sturm 2002).

A concerted effort is required to employ standardized equipment, formalize techniques and methodologies and extend the temporal record of the NRB through additional fieldwork. Pressure should be placed on governments to enhance weather station networks instead of diminishing them and to be diligent in the correction of both snow and rainfall records. The sparse and declining number of weather stations collecting precipitation data throughout the circumpolar north also makes the development of remote sensing technology, which is cost effective, reliable and has frequent coverage more vital, especially in the context of a changing climate.

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