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Evidence of meltwater retention within the Greenland ice sheet

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Abstract

Greenland ice sheet mass losses have increased in recent decades with approximately half of these attributed to increased surface meltwater runoff. However, controls on ice sheet water release, and the magnitude of englacial storage, firn densification, internal refreezing and other hydrologic processes that delay or reduce true water export to the global ocean remain poorly understood. This problem is amplified by scant hydrometeorological measurements. Here, ice sheet surface meltwater runoff and proglacial river discharge determined between 2008 and 2010 for three sites near Kangerlussuaq, western Greenland were used to establish the water budget for a small ice sheet watershed. The water budget could not be closed in the three years, even when uncertainty ranges were considered. Instead between 12 % and 53 % of ice sheet surface runoff is retained within the glacier each melt year (time between onset of ice sheet runoff in two consecutive years). Evidence of the ice sheet summer meltwater escaping during the cold-season suggests that the Greenland ice sheet cryo-hydrologic system may remain active year round.

1 Introduction

Greenland ice sheet mass losses from ice discharge and meltwater runoff have almost tripled since 1958 (Rignot et al., 2008). Recent meltwater runoff losses comprise a substantial fraction of total losses, with cumulative meltwater runoff anomalies estimated to be twice as large as cumulative ice discharge anomalies between 2000 and 2008 (van den Broeke et al., 2009). This meltwater loss intensification is consistent with observations of rising mean annual near-surface air temperature (+1.8 °C between 1840 and 2007, Box et al., 2009) and expanding melt area on the ice sheet surface (Abdalati and Steffen, 2001; Mote, 2007; Tedesco, 2007), including a record high in 2010 of approximately double the 1979–2009 average (Tedesco et al., 2011). Furthermore, reduced ice sheet albedo associated with surface melting amplifies its sensitivity to increasing

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air temperature (Box et al., 2012), suggesting even further importance of meltwater runoff losses.

Continued mass loss from the Greenland ice sheet has the potential to raise global sea levels by 8 ± 4 cm by 2050 (Rignot et al., 2011) and between 17 and 54 cm by 2100 (Pfeffer et al., 2008). However, such estimates assume unimpeded evacuation of meltwater from the ice sheet surface to the ocean. It has long been known that meltwater transport from the ice sheet surface to its margin occurs through a complex, poorly understood system of supra-, en- and subglacial pathways consisting of crevasses, moulins, fractures, conduits and supraglacial stream channels (Fountain and Walder, 1998), collectively referred to as the cryo-hydrologic system (CHS) (Phillips et al., 2010). Meltwater passage through the CHS and temporary storage(s) in supraglacial lakes and englacial cavities can retard gravity-driven flow of melt water from its creation on the ice sheet surface to its appearance at the ice margin (Cuffey and Paterson, 2010; Fountain and Walder, 1998). Furthermore, meltwater can be retained if it refreezes or accumulated in firn layers (Boggild, 2007; Boggild et al., 2005; Fausto et al., 2009; Greuell and Konzelmann, 1994; Pfeffer et al., 2008; Reeh, 1991).

CHS transport, storage, and retention are known to cause significant delays in ice sheet and glacier meltwater release. Internal glacier water storage can redistribute water losses seasonally by collecting water in the early melting-season, and releasing it in the late melting-season (Jansson et al., 2003) or in winter months (Hagen et al., 2003; Hodson, 2005; Jansson et al., 2003; Stenborg, 1965; Wadham, 2000). Additionally, sudden drainage of ice sheet supra- (Bartholomew et al., 2011b), en-, sub- (Mathews, 1963), and pro-glacial meltwater lakes and storages (Mernild and Hasholt, 2009; Mernild, 2009; Russell, 2009; Russell et al., 2011) can result in pronounced river discharge anomalies.

While such delays and/or reductions of melt water fluxes to the global ocean by englacial processes are widely appreciated, they remain poorly quantified. Direct observations of ice sheet meltwater runoff extending over multiple years (van den Broeke et al., 2011) and ice sheet runoff losses through river discharge (Ahlstrøm et al., 2002;

Mernild and Hasholt, 2009; Rennermalm et al., 2012) are scarce for Greenland. Instead, these losses are calculated from satellite gravity anomalies (Chen et al., 2011; Luthcke et al., 2006; Ramillien et al., 2006; Velicogna and Wahr, 2005), remotely-sensed elevation changes at the ice sheet surface (Krabill et al., 2004; Pritchard et al., 2009) or inferred from surface mass balance models (Box et al., 2006; van den Broeke et al., 2009; Ettema et al., 2009; Fettweis, 2007; Hanna et al., 2008; Mernild et al., 2009, 2010a). All of these methods have uncertainties when used to estimate meltwater fluxes exiting the ice sheet. For example gravity and altimetry measurements cannot separate dynamic losses from meltwater losses without relying on surface mass balance models (Krabill et al., 2004) or assumptions about geographic distribution of dominant mass loss processes (Pritchard et al., 2009). Runoff estimates inferred from surface mass balance models vary widely, ranging between 248 and 407 $km^3 yr^-$ (Box et al., 2006; Ettema et al., 2009; Fettweis, 2007; Hanna et al., 2008; Mernild et al., 2009, 2010a). This variability reflects large uncertainties owing to model biases, initial conditions, and differences in how models resolve hydrologic processes such as meltwater refreezing (Fettweis, 2007). Additional uncertainty arises from choice of model resolution (Ettema et al., 2009; Fettweis, 2007), representativeness of the spatial distribution of model forcing data (Box et al., 2006), and ignoring water storage for example in supraglacial lakes, which are abundant on the Greenland ice sheet (Selmes et al., 2011).

Here, the absolute volumes of meltwater runoff on the ice sheet surface, and its appearance in proglacial rivers as discharge by the ice margin, are compared to quantify englacial storages and transport near Kangerlussuaq, west-central Greenland using two recently published datasets (van den Broeke et al., 2011; Rennermalm et al., 2012). The first provides ice sheet runoff estimated from a surface energy balance model using data from two automatic weather stations (AWS) along the K-transect (van den Broeke et al., 2011). The K-transect is an array of eight surface mass balance observation points operational since 1990, that extend 141 km into the ice sheet interior from the Russell Glacier terminus in Southwest Greenland (van de Wal et al., 2005).

The ice sheet surface energy balance model factor in meltwater retention, thus providing estimates of ice sheet runoff input into the ice sheet watershed CHS. The second dataset provides proglacial river discharges since 2008, measured 2 km from the ice margin in the Akuliarusiarsuup Kuua (AK) River, just north of Russell Glacier (Rennermalm et al., 2012). While ice sheet runoff provides CHS meltwater input, river discharge provides estimates of meltwater output exiting the CHS and escaping to streams and lakes by the ice sheet margin so that a water budget for the ice sheet watershed can be constructed. These two terms (CHS input and output) are subtracted to determine ice sheet meltwater release and retention both cumulative over each melt-year (time between ice sheet runoff onset in two consecutive years), and continuously. To increase confidence in cold-season meltwater release events at times with limited or no ice sheet surface melting, in-stream temperature records are analyzed for ice to water phase change evidence.

2 Study site

The study watershed is situated along the Greenland ice sheet's southwestern margin ~30 km northeast of Kangerlussuaq, between the Isunnguata Sermia and Russell outlet glaciers (Fig. 1). Its area is 64.2 km^2 and was delineated using flow directions derived from the ASTER GDEM surface elevation dataset (ASTER GDEM Validation team, 2009). While incorporation of high-resolution basal topography data would enable more accurate hydraulic drainage delineation using the hydrostatic potentiometric surface (Cuffey and Paterson, 2010; Lewis and Smith, 2011), the use of ice surface elevations alone to identify hydraulic flow directions and watershed boundaries is common (Mernild et al., 2010b) despite these limitations (van As et al., 2012). The study watershed is mostly ice-covered (60.0 km^2) with a small proglacial area (4.2 km^2) consisting of exposed bedrock, tundra, loess deposits, lakes, and river alluvium. The ice sheet part of the study watershed is hereafter called AK4 ice sheet watershed. Ice sheet surface elevations within this AK4 ice watershed range from 500 to 860 m a.s.l.

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Since 2008, hydrometerological observations have been collected at three sites along the Akuliarusiarsuup Kuua River (Rennermalm et al., 2012) as follows: (1) river discharge, stream water level and stream temperatures (Site AK4, Fig. 1); (2) lake stage fluctuations (Site AK5, Fig. 1); (3) and near-surface air temperatures (Site AK1 and AK2, Fig. 1). Meteorological observations on the ice sheet were acquired at three Automated Weather Stations (AWS) labeled S5, S6, and S9 on Fig. 1 at 490, 1020 and 1520 m a.s.l., respectively (van den Broeke et al., 2011), as well as Kangerlussuaq airport (~ 30 km west southwest of AK4 on Fig. 1).

3 Methods

River discharge was determined by relating discharge measurements with half-hourly in-stream pressure recordings corrected for background atmospheric pressure variability (Rennermalm et al., 2012). Data were collected with high-precision Price Type-AA current meters, and Solinst Level- and Baro-loggers. River discharge time series quality was assessed, and the ~ 68 % confidence interval determined to be 18 % of discharge (Rennermalm et al., 2012). At Site AK4 river discharge was measured between 9 June 2008 and 17 August 2010. At Site AK5 water levels were measured between June 2007 and 19 August 2008. To include the start of the 2008 melting season, AK4 river discharge time series was retroactively estimated to January 2008 through regression with AK5 water levels using data from an overlapping period between 9 June and 19 August 2008. This relationship was described with a power law ($Q_{\text{AK4}} = 2.2\,L_{\text{AK5}}^{0.70}$ where Q_{AK4} is AK4 river discharge, and L_{AK5} is AK5 water level) explaining 70 % of river discharge variability observed at AK4.

Ice sheet runoff volume at the point locations of Sites S5, S6, and S9 was determined using a surface energy balance model relying on data inputs of surface momentum roughness, snow depth, wind speed, temperature, humidity, down- and upwelling shortwave radiation, and downwelling longwave from automatic weather stations and sonic depth rangers (van den Broeke et al., 2011, 2004, 2008a, b, 2009b), and factoring in percolation and refreezing in snow layers using the Greuell and Konzelmann (1994) methodology. S5, S6, and S9 AWS stations have been operational since 1 September 2003, except for an S6 datalogger failure between 2 September 2007 to 3 September 2008 and an S6 data gap between 13 July to 17 August 2010. During overlapping data periods, S5 data captured 73 % of daily S6 runoff variability with a linear regression model ($R_{S6} = 0.59 - 2.6 R_{S5}$, where R_{S6} and R_{S5} is runoff at S5 and S6 respectively). Using this relationship, aforementioned S6 data gaps were filled except between 19 April and 27 May 2008 when S6 was forced to zero to reflect a 38 day average delay in runoff onset at S6 relative to S5.

Daily runoff volume ($m^3\,d^{-1}$) from AK4 ice sheet watershed (R_W) was determined by constructing a linear elevation-dependent runoff model using inputs from S5 and S6:

$$
\begin{aligned}
R_{Wi} &= A_{AK4} \sum_{j=1}^{N_B} (a_i E_j + b_i) f_j \\
a_i &= (R_{S5} - R_{S6}) / (E_{S5} - E_{S6}) \\
b_i &= R_{S5} - a_i E_{S5}
\end{aligned}
\tag{1}
$$

R_{Wi} is watershed runoff volume on day i ($m^3\,d^{-1}$), A_{AK4} is ice sheet watershed area upstream AK4 (m^2), a_i and b_i are intercept and slope on day i ($md^{-1}\,m^{-1}$), E_j is elevation bin midpoint value for elevation band j (m), f_j is elevation band fractional area, N_B is number of elevation bands. Slope and intercept were determined using daily ice sheet runoff at S5 (R_{S5}) and S6 (R_{S6}), and their respective elevation (E_{S5} and E_{S6}). Given that S5 or S6 are located just outside the AK4 ice watershed, this model implicitly assumes that elevation is a dominant control on ice sheet runoff. Indeed, glacier surface melt models often divide watersheds into elevation bands (Hock, 2005) to represent temperature lapse rate, and inverse proportionality between elevation and net radiation (van den Broeke et al., 2011). Model validation was made by assessing elevation dependent meltwater production using Eq. (1) (runoff volume is substituted with meltwater production), which in contrast to runoff occurs at all three ice sheet AWS stations.

This elevation dependent model parameterized with meltwater runoff production at S5 and S6 captured 91 % of daily S6 meltwater production ($r = 0.96$, $p = 0.01$). In contrast to meltwater production, ice sheet runoff estimates factors in meltwater retention and can therefore be assumed representative of the amount of ice sheet meltwater that escape to the rivers in the pro-glacial environment (e.g. AK4). Recognizing ice sheet watershed delineation uncertainties, an upper and lower range of ice sheet runoff was determined as $\pm 0.1\,R_W$ corresponding to the weight of basal topography in the potentiometric surface calculations (Cuffey and Paterson, 2010).

Characterization of meteorological conditions on the ice sheet and in the pro-glacial tundra environment were examined with near-surface air temperatures obtained from the Kangerlussuaq AWS station 042310 (National Climatic Data Center, 2011), S5 AWS station (S5 was chosen over S6 due its uninterrupted time series), and AK1 and AK2 Solinst Barologgers©. Barologger air temperatures determined with unshielded sensors were suitable for examining winter/early spring conditions when the sun has limited sensor interference due to low solar angles and low radiation. Snow depth was retrieved from the S5 AWS station.

4 Results

Cumulative discharge in the AK4 River (Q_{AK4}) is always less than upstream ice sheet watershed runoff volumes (R_W) (Fig. 2a). Right before ice sheet spring melt onset when the cumulative runoff reaches its maximum at each melt-year end date, cumulative R_W was 112 %, 178 %, and 169 % of Q_{AK4} in 2008/09, 2009/10, 2010/11. Thus, all years exhibit river discharge deficits, meaning that more runoff is produced at the ice sheet surface than is observed in the downstream proglacial river channel. Although the upper range of Q_{AK4} confidence interval and lower range of R_W close the ice sheet watershed budgets and reduce ice sheet meltwater retention in 2008/09, the watershed budget gap and retention is still considerable in 2009/10 and 2010/11. Ice sheet runoff that does not escape to the stream might instead be retained in glacial and proglacial

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storages. The fraction retained ice sheet runoff increase throughout each melt year (Fig. 2b) and add up to 12 %, 49 %, and 53 % at each melt-year end date in 2008/09, 2009/10, 2010/11.

Daily summer river discharge and ice sheet runoff (15 June to 15 August) co-vary strongly (Fig. 3), with correlation coefficients between 0.74–0.91. While river discharge is less compared to daily ice sheet meltwater runoff volume, their relationship is linear and similar in all years.

Indeed, daily river discharge generally covaries with ice sheet runoff but with a lower magnitude and a dampened signal (Fig. 4a). Exceptions to this pattern are cold-season river discharge events, which precedes ice sheet spring onset in 2008 and 2009, and results in lengthy periods of river discharge surplus at those times possibly due to either ice sheet meltwater release or land-based snow melt in the proglacial environment (river surplus is shown as negative values in Fig. 4b). In 2008 and 2009, river flow commenced within 0–2 days of isolated ice sheet runoff pulses, but was followed by 18–23 days without significant ice sheet runoff (Table 1). In 2010 river discharge lagged ice sheet runoff onset by 2 days. In 2008, river discharge onset is 31 days earlier than thaw onset at S5, but coincides within 3 days of thaw at Kangerlussuaq (Table 1). In 2009 and 2010, opposite patterns are detected when river discharge onset coincides with thaw at S5 (within 1–2 days), but is considerably delayed compared to thaw at Kangerlussuaq (32–35 days).

Three prolonged periods of river flow occur at times with no significant ice sheet meltwater runoff production, which is suggestive of cold-season release of englacial meltwater. One period occurs in winter of 2008 between 31 October and 28 November. The two others occur in pre-melting season months (March and April) between 23 March and 16 April in 2008 and between 27 April and 15 May in 2009 (Fig. 5, middle panel). The 2008 winter runoff event is identified as a water pressure anomaly between 30 October and 28 November. It is accompanied by a rapid increase in stream temperature from −5.5 °C to −0.5 °C between 26 and 28 October, followed by a period of constant near-zero (average is −0.53 °C) stream temperature until 28 November

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(Fig. 5) indicative of water phase change from frozen to liquid. This river discharge anomaly is preceded by a small ice sheet runoff pulse between 27 and 30 October 2008, coinciding with above freezing near surface air temperatures at both AK2 and S5 (Fig. 5). River discharge peaks on 5 November, seven days after runoff peaked on the ice sheet. Although peak water flow is lower in the stream compared to the ice sheet, cumulative river discharge between 28 October and 29 November exceeds ice sheet runoff by 6 times (14×10^6 m^3 vs. 2.3×10^6 m^3).

The two other river discharge events observed in March/April 2008 and April/May in 2009 are also registered as pressure anomalies (Figs. 6 and 7). In March/April 2008 discharge was inferred from a sensor at AK5 installed at a lake bottom and cannot be used to determine stream temperatures at AK4 during the anomaly. In contrast, April/May 2009 discharge was determined with a sensor at AK4. In 2009, stream temperature increase by 5 °C on 27 April in concert with rising water pressure. Although AK4 stream temperatures remain below freezing, this may be indicative of a shallow layer of thawing ice underneath flowing water. Similar to the 2008 October/November discharge event, these two pre-melting season events were preceded by small ice sheet runoff pulses (22–23 March 2008 and 25 April 2009), and short periods (hours) of above zero air temperatures on the ice sheet and in the proglacial environment (Figs. 6 and 7). At the time of their release, cumulative annual river discharge volumes for these events were 22 and 56 times larger than cumulative ice sheet meltwater runoff over the same time period.

Although precipitation or snow melt, and/or measurement uncertainties and errors cannot be ruled out, cold-season release of englacially stored meltwater seems the most likely explanation for the in-stream pressure increases recorded during winter 2008 (31 October and 28 November) and early spring seasons of 2008 and 2009 (23 March–16 April and 27 April–15 May, respectively, Figs. 5, 6, and 7). These pressure anomalies correspond to maximum water depths of 0.6 m, 0.3 m, and 0.6 m respectively, which is well within the typical range of flow depths observed at Site AK4 (0–1.4 m). Snow accumulation seems a less likely explanation, requiring depths of

3.2–6.3 m and 0.54–1.3 m, respectively 10 % and 50 % of water density. Such depths are markedly higher than typical snow accumulation at the S5 and S6 AWS sites of 0.16 m and 0.33 m respectively.

5 Discussion

This study finds that a large fraction of ice sheet surface runoff produced each melt-year (time between ice sheet runoff onset in two consecutive years) is retained within the ice sheet watershed (12 %, 48 %, and 53 % in 2008/09, 2009/10, and 2010/11). While some retained meltwater is delayed and released in the cold-season, it does not close the gap between ice sheet watershed meltwater inputs and outputs even when uncertainty ranges for runoff and discharge are considered. It is known that Greenland ice sheet meltwater may be retained on the ice sheet surface in lakes (Selmes et al., 2011; Sundal et al., 2009) and percolate into firn layers (Humphrey et al., 2012). However, this study suggests that meltwater transported through englacial conduits (Catania and Neumann, 2010; Catania et al., 2008) to storage cavities in the subglacial environment may remain there until short thaw events trigger its release. Such subglacial storages, and delayed release, have been observed for high Arctic glaciers in Svalbard (Hodson, 2005; Wadham et al., 2001), but not previously identified for the Greenland ice sheet.

Evidence for delayed ice sheet surface meltwater release is found in daily, seasonal and longer time scales. First, ice sheet surface runoff and downstream river discharge have high co-variability (0.74–0.91) in summer months, but the river discharge signal is dampened. This is typical of the impact of meltwater transport in supra-, en-, and sub-glacial environments during its passage to rivers at the ice terminus (Fountain and Walder, 1998). Second, three considerable cold-season meltwater releases during times with insignificant ice sheet surface melting suggest that the CHS system can retain meltwater for at least 1 to 6 months after surface melting season ends in September. Finally, the water balance between ice sheet surface runoff and river discharge

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results in net meltwater storage three years in a row suggest meltwater retention on timescales beyond one year. Delays on monthly and longer time scales are not unique to the Greenland ice sheet, but have been identified for other Arctic glaciers (Hagen et al., 2003; Hodson, 2005; Jansson et al., 2003; Stenborg, 1965; Wadham, 2000).

The three cold-season meltwater releases were only accompanied by modest ice sheet surface runoff, but two pieces of evidence show that river channel flow indeed did occur between 31 October to 28 November 2008, 23 March to 16 April 2008, and 27 April to 15 May 2009. First, in-stream pressure anomalies during events cannot be explained by sudden dense snow packs; instead, derived water levels and river discharge are well within the natural range at AK4. Second, in-stream temperatures suggest liquid water, and/or phase change. During October 2008, in-stream temperatures co-vary with near surface air temperatures until 31 October because there is no water evident at AK4 (Fig. 5). However, after November 1st in-stream temperatures rise, air temperatures remain below freezing, coinciding with in-stream pressure increase indicating flowing unfrozen water. During the 2009 pre-melt season, a sudden 5 °C in-stream temperature increase coincided with raised in-stream pressure indicate melting and flowing water (Fig. 6). In the pre-melting season of 2008, meltwater release precedes ice sheet surface melting and above-freezing near-surface air temperatures (Fig. 7). Thus, despite inherent uncertainties of wintertime river low-flow observations (Pelletier, 1990), existence of cold-season ice sheet meltwater discharge is evident. Measurement uncertainties and errors are unlikely given that the sensor operated as expected before and after the three river runoff events, and coherence with brief preceding ice sheet runoff events points toward broader scale events. Presence and release of unfrozen meltwater in the Greenland ice sheet at sub-freezing temperatures suggest that parts of the cryo-hydrologic system are intact in the cold-season. Indeed, this possibility is confirmed by modeling studies (Phillips et al., 2010), and observations with ground penetrating radar (Catania and Neumann, 2010).

Above-freezing air temperatures at the time of the cold-season meltwater pulses may provide a triggering mechanism for release of stored meltwater. During these times, a short thaw period (1–4 days) started at the S5 AWS site 30–75 h before they were detected in the stream. Although bursts of ice sheet runoff accompany all events, their duration and magnitude are too short (1–3 days vs. 18–30 days) and too small (2–16 % of river discharge volume in cold-season pulses) to explain all river runoff observed during these periods. Lack of significant ice sheet runoff during these months indicates that cold-season river discharge was produced during warm summer months. Additional drivers for winter release may stem from buildup of subglacial pressure as englacial meltwater drains to subglacial cavities and drainage exits close (Irvine-fynn et al., 2011).

Cold-season release may in fact be a consequence of meltwater retention within the Greenland ice sheet according to the following hypothesis: meltwater retention builds up subglacial pressures that maintain a largely intact CHS in the cold-season, so that the CHS can be readily activated and allow meltwater release in response to triggers such as short lived thaw events. This hypothesis is not unique to this study. Six years of energy and mass balance studies from a Svalbard glacier reveal that years with lower than expected ice sheet meltwater export (due to internal storage) preceded years when less than usual energy was required to open the subglacial system, perhaps due to increased subglacial pressure from larger than usual internal storage (Hodson, 2005). It is unclear how meltwater retention influences CHS evolution. However, modeling and observational studies suggest that CHS seasonal evolution from an un-channelized steady state with small conduits (i.e. cavity dominated) to an efficient channelized system with large conduits is controlled by meltwater supply rates to the subglacial hydrologic drainage system (Schoof, 2010; Sundal et al., 2011). Furthermore, in years with strong ice sheet surface melting, and thus more efficient sub-glacial drainage, ice sheet velocities decelerate early (Sundal et al., 2011). Following the logic that a faster flowing ice sheet can enhance surface melting when more ice is brought to higher temperatures at lower elevations (Parizek and Alley, 2004), the subglacial drainage system's effectiveness (Bartholomew et al., 2011a; Sundal et al., 2011) becomes an additional control to mass balance (Hanna et al., 2008; Mernild et al., 2010a)

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and future meltwater losses from the Greenland ice sheet's land-terminating glaciers. Besides the potential links between meltwater retention, CHS, and dynamic ice losses, retained meltwater also delays ice sheet water losses to the ocean. To understand the importance of Greenland ice sheet meltwater retention on these processes, more observational and modeling studies must establish how common this is in other parts of Greenland, and how it is related to subglacial pressure, cold-season releases, and CHS development and functioning.

6 Conclusions

Greater cumulative ice sheet surface runoff relative to downstream river discharge suggests meltwater retention in en- and subglacial storages. Indeed, three observed river runoff events outside of the regular summer melting period, and preceded by very small ice sheet runoff fluxes, could be examples of delayed release of en- and sub-glacially stored water. Thus, parts of the Greenland ice sheet CHS may remain active year round. More information is needed to determine how important to meltwater retention is to development, and functioning of the melting season CHS. Further investigations are needed to establish how widespread meltwater retention and delayed releases are along the Greenland ice sheet perimeter.

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Table 1. Timing of onset of river discharge, ice sheet surface runoff, and thaw proxy (dates when cumulative near surface air temperature exceed 0 °C at S5 and Kangerlussuaq AWS). Dates within parentheses shows isolated melting events before melting season onset.

Site	2008	2009	2010
Flow onset			
River discharge (AK4)	23 Mar	27 Apr	3 May
Ice sheet watershed runoff	16 Apr (22–23 Mar)	15 May (25 Apr)	1 May
Thaw onset			
S5 (ice sheet)	23 Apr	26 Apr	1 May
Kangerlussuaq (pro-glacial)	26 Mar	5 Mar	26 Mar

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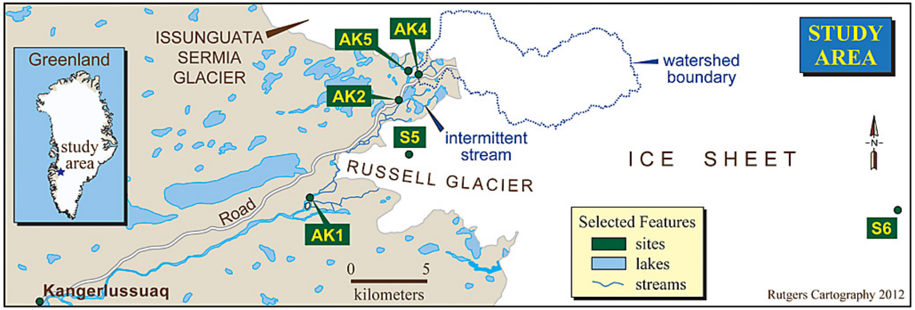

Fig. 1. Map of study area showing monitoring installations for proglacial river discharge (AK4), and its estimated watershed boundary (hatched area), nearby lake-level (AK5), and air temperature (AK1 and AK2) of Rennermalm et al. (2012). Supraglacial AWS monitoring sites S5 and S6 of van den Broeke et al. (2011) are also shown, the S9 AWS station is located 82 km east of AK4. An additional AWS station operated by the Danish Meteorological Institute is located in Kangerlussuaq.

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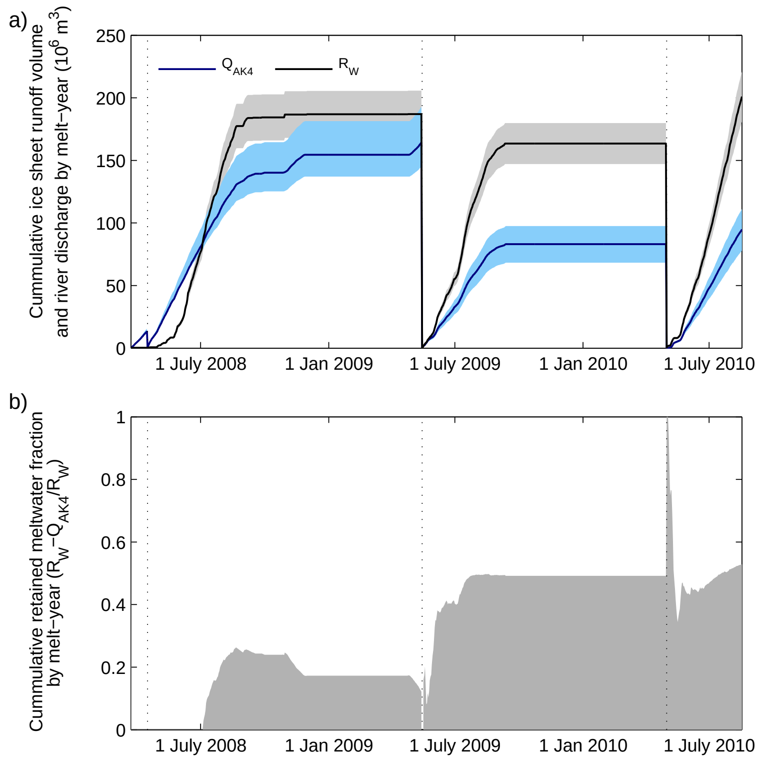

Fig. 2. (a) Cumulative discharge at the AK4 river site, Q_{AK4} and cumulative runoff volume from its upstream ice sheet watershed, R_W, from 23 March 2008 to 17 August 2010 by melt-year (i.e. reset at ice sheet spring runoff onset each year). Grey shaded regions show $R_W \pm 0.1\,R_W$, blue shaded regions show upper and lower confidence interval boundaries of Q_{AK4}; **(b)** cumulative retained meltwater fraction by melt-year determined by subtracting cumulative Q_{K4} from cumulative R_W and dividing by cumulative R_W.

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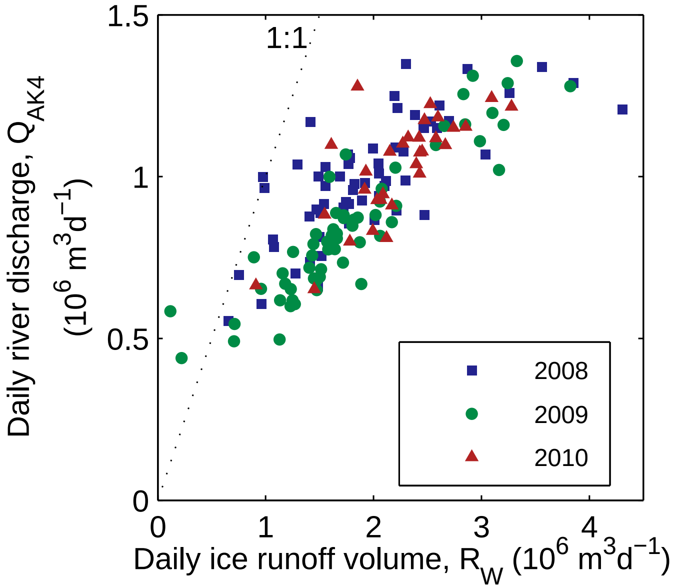

Fig. 3. Daily river discharge, Q_{AK4} as a function of ice sheet runoff volume, R_W between 15 July and 15 August in 2008, 2009 and 2010. While ice sheet runoff volume is larger than river discharge (1 : 1 line shown as stippled line), the two variables have strong linear relationships in all years with correlation coefficients between 0.74 and 0.91.

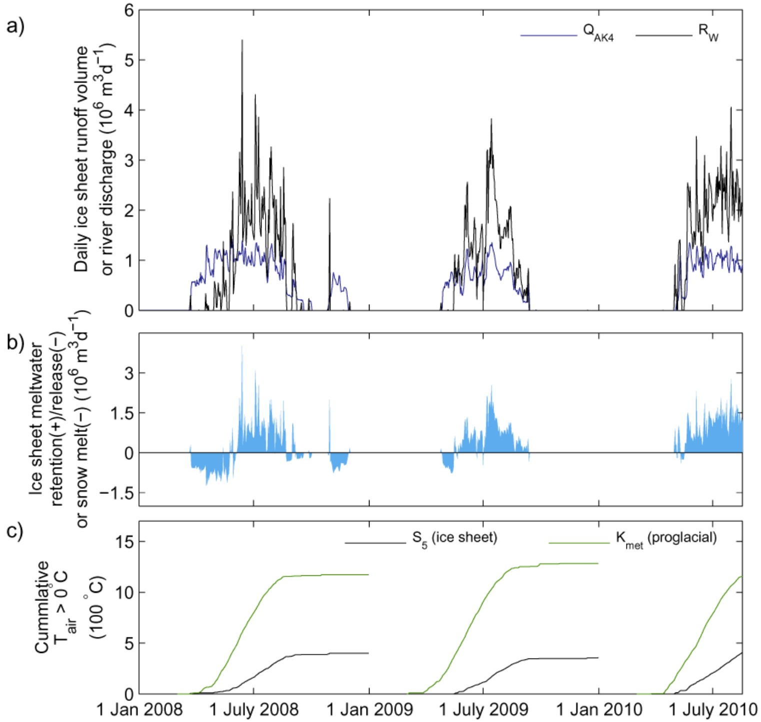

Fig. 4. Daily river discharge and ice sheet runoff volume between 1 January 2008 and 17 August 2010 **(a)**, ice sheet meltwater retention/ejection **(b)**, and cumulative daily near-surface air temperature (T_{air}) above 0 °C after 1 March each year at S5 and Kangerlussuaq AWS **(c)**. River discharge starts earlier in 2008 and 2009 and has a dampened diurnal variability compared to ice sheet runoff. The date when cumulative T_{air} pass zero, a proxy for thaw, indicate earlier melt season onset and longer duration in proglacial areas (Kangerlussuaq) than on the ice sheet (S5).

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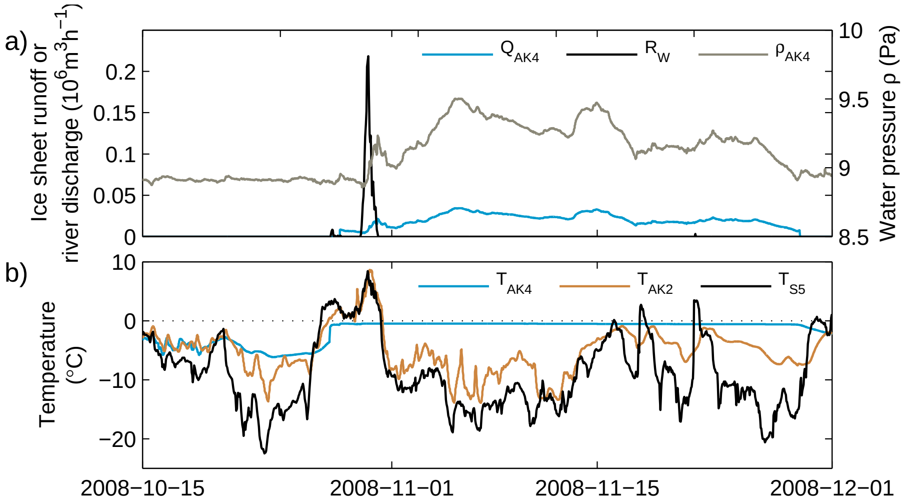

Fig. 5. Ice sheet, river, and pro-glacial conditions during the October–November 2008 runoff event, including discharge, runoff and sensor water pressure **(a)**, and temperatures **(b)**. A short ice sheet runoff pulse (R_W) accompanied above zero ice sheet (T_{S5}) and proglacial (T_{AK2}) surface air temperatures precedes a longer river discharge event (Q_{AK4}). Comparison of surface air and in-stream temperatures reveals when the latter is exposed to air (15–28 October), and when it is not (28 October–1 December). On 1 November, in-stream pressure increases above background noise and in-stream temperatures rise rapidly to near-zero suggesting that the AK4 in-stream sensor is submerged by liquid water after this date.

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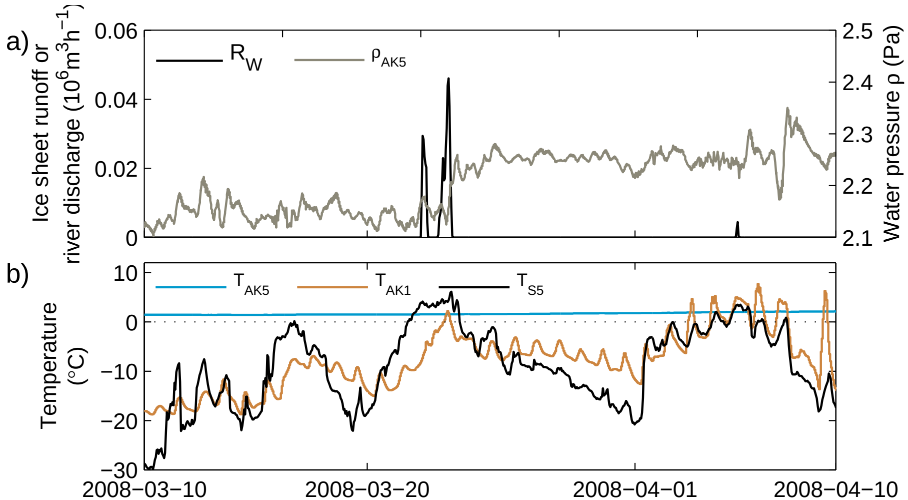

Fig. 6. Ice sheet, river, and pro-glacial conditions during the 2008 pre-melting season runoff event, including runoff, discharge and sensor water pressure **(a)**, and temperatures **(b)**. A short ice sheet runoff pulse between 22 and 23 March (R_W) precedes a marked increase in water pressure at AK5 (ρ_{AK5}, AK4 Sites was not operational at this time and extrapolated values for this site are not shown). The ice sheet runoff between 22 and 23 March coincide with a short period with above zero ice sheet (T_{S5}) and proglacial temperatures (T_{AK1}). Constant lake temperatures (T_{AK5}) during this period indicate unfrozen conditions at the AK5 lake bottom.

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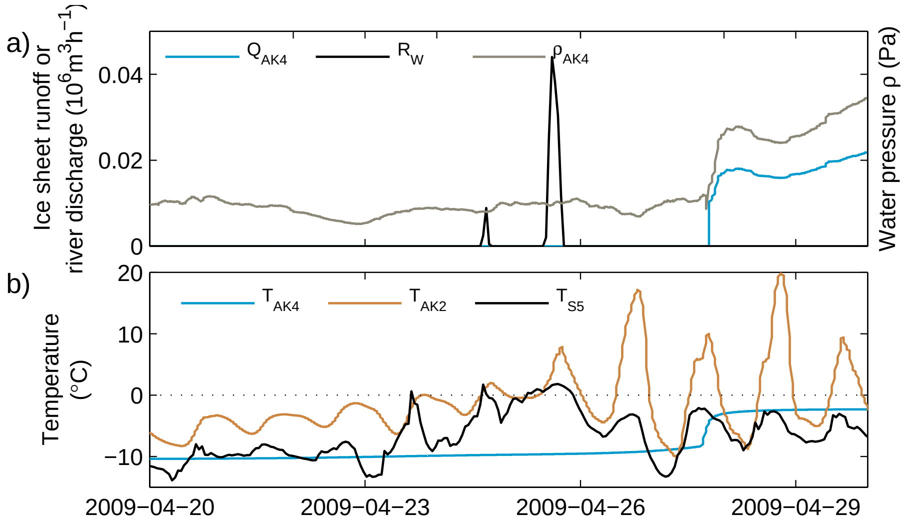

Fig. 7. Ice sheet, river, and proglacial conditions during the 2009 pre-melting season runoff event, including runoff, discharge, and sensor water pressure **(a)**, and temperatures **(b)**. A short ice sheet runoff pulse (R_W) on 25 April accompanied above zero ice sheet (T_{S5}) and proglacial (T_{AK2}) surface air temperatures precedes river runoff onset with 2 days (Q_{AK4}). In-stream pressure on 27 April increases simultaneous with a 5 °C rise of in-stream temperatures, which are suggestive of phase change and liquid water flow after this date.

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