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Key Points:
- We present a high-resolution subglacial hydrological analysis of the Greenland ice sheet
- Small subglacial lakes remain undetectable by methods using surface elevation or radar techniques
- We identify the first evidence for subglacial water piracy beneath the Greenland ice sheet

Supporting Information:
- Supporting Information S1

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Subglacial water drainage, storage, and piracy beneath the Greenland ice sheet

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Abstract Meltwater drainage across the surface of the Greenland ice sheet (GrIS) is well constrained by measurements and modeling, yet despite its critical role, knowledge of its transit through the subglacial environment remains limited. Here we present a subglacial hydrological analysis of a land-terminating sector of the GrIS at unprecedented resolution that predicts the routing of surface-derived meltwater once it has entered the basal drainage system. Our analysis indicates the probable existence of small subglacial lakes that remain undetectable by methods using surface elevation change or radar techniques. Furthermore, the analysis suggests transient behavior with rapid switching of subglacial drainage between competing catchments driven by seasonal changes in the basal water pressure. Our findings provide a cautionary note that should be considered in studies that attempt to relate and infer future response from surface temperature, melt, and runoff from point measurements and/or modeling with measurements of proglacial discharge and ice dynamics.

1. Introduction

Mass loss from the Greenland ice sheet (GrIS) accelerated over the two last decades, and the GrIS is expected to be the largest cryospheric contributor to global sea level rise in the 21st century [*Hanna et al.*, 2013]. To predict the future contribution of the GrIS to sea level rise, we must understand the physical processes that govern ice sheet mass balance and dynamics. Despite this, the important role that subglacial water routing and storage plays in ice sheet dynamics remains poorly constrained. Several recent studies [e.g., *Bartholomew et al.*, 2010] have argued that hydrological processes on valley glacier systems could be scaled up to serve as analogues for ice sheets. While this analogue appears to hold for the ice sheet margin [e.g., *Sundal et al.*, 2011], the considerable differences in geometry between valley glaciers and ice sheets become pronounced further inland, where the development of efficient subglacial hydrology is hindered by low surface melt rates and gentle bed slopes [*Meierbachtol et al.*, 2013; *Rennermalm et al.*, 2013b; *Dow et al.*, 2014; *Doyle et al.*, 2014]. Direct observations of the basal hydrological system from boreholes or tracing experiments are limited to the lower ablation zone of the GrIS [*Smeets et al.*, 2012; *Chandler et al.*, 2013; *Meierbachtol et al.*, 2013; *Andrews et al.*, 2014; *Doyle et al.*, 2015; *van de Wal et al.*, 2015] while geophysical data covering the whole ice sheet (used for subglacial water routing) [e.g., *Livingstone et al.*, 2013] have only been available at a low spatial resolution (1 km grids of bed elevation data) [*Bamber et al.*, 2013a]. Nevertheless, several studies, which were based on a combination of high-resolution ice surface and low-resolution bed digital elevation models (DEMs), have concluded that there is substantial subglacial or englacial meltwater storage [*Rennermalm et al.*, 2013a; *Smith et al.*, 2015]. The lack of high-resolution bed topography data has also led previous comparative studies of ice surface runoff and proglacial discharge to extrapolate subglacial drainage catchments from ice surface DEMs [e.g., *van de Wal and Russell*, 1994; *Mernild and Hasholt*, 2009; *Mernild et al.*, 2010; *Bartholomew et al.*, 2011; *Palmer et al.*, 2011; *van As et al.*, 2012; *Chandler et al.*, 2013; *Cowton et al.*, 2013; *Fitzpatrick et al.*, 2014]. Yet when water leaves the supraglacial system, its pathway and rates of flow to the ice sheet margin is largely unknown and the water does not necessarily follow the ice surface topography.

Hydraulic potential analysis enables subglacial drainage to be estimated from ice thickness and basal topography on spatial scales where the main control of the drainage is the geometry of the ice sheet [*Shreve*, 1972]. This method has been applied in Greenland [e.g., *Lewis and Smith*, 2009; *Bamber et al.*, 2013b; *Banwell et al.*, 2013;

Livingstone et al., 2013; *Karlsson and Dahl-Jensen*, 2015] and Antarctica [e.g., *Wingham et al.*, 2006; *Wright and Siegert*, 2012; *Wolovick et al.*, 2013] to study subglacial water flow and to identify the locations of subglacial lakes. Antarctic subglacial hydrological flow paths have shown to be highly sensitive to changes in ice surface elevation and may also exhibit highly unstable conditions [*Fricker et al.*, 2007; *Wright et al.*, 2008; *Allison et al.*, 2009] depending on the subglacial water pressure regime. In some cases, ice flow has been observed to switch on and off owing to water competition between adjacent ice flow units, a behavior termed water piracy [*Anandakrishnan and Alley*, 1997; *Vaughan et al.*, 2008; *Carter et al.*, 2013]. Water piracy has not yet been considered in Greenland; this is surprising given its important role in marginal ice flow dynamics. Variations in subglacial water pressures are often large in Greenland and are caused by the seasonal evolution of the subglacial drainage system [*Meierbachtol et al.*, 2013; *Andrews et al.*, 2014; *van de Wal et al.*, 2015] and short-term variations in surface melt runoff [*van de Wal et al.*, 2008; *Bartholomew et al.*, 2012]. Hence, we expect that water piracy may also have an effect on subglacial water routing beneath the GrIS. Here we apply hydraulic potential analysis to the Kangerlussuaq sector of the GrIS. We use high-resolution DEMs to (1) compare modeled ice surface runoff with measured proglacial discharge and (2) test the hypothesis that subglacial water piracy exerts an important control on GrIS dynamics and proglacial discharge.

2. Data and Methods

We applied hydraulic potential calculations to a ~ 8500 km^2 land-terminating area of the Kangerlussuaq sector of the GrIS to investigate subglacial water routing and storage. Hydraulic potential is a steady state proxy for routing of subglacial water [*Shreve*, 1972] and is the sum of the pressure potential from the overburden ice and the elevation potential. We used a 250 m resolution ice thickness DEM from the compiled radar data sets of *Lindbäck et al.* [2014], supplemented with the mass conservation model of *Morlighem et al.* [2014] at 150 m resolution. We calculated the bed elevation from the latter by subtracting the ice thickness from the GrIS Mapping Project (GIMP) surface elevation model at 30 m resolution [*Howat et al.*, 2014]. To capture sufficient topographic detail and accuracy for our analysis, we used a DEM with a resolution of 150 m, which represents a compromise between the differing horizontal resolutions of the data sets. We used a single-direction flow algorithm and surface analysis (see Text S1 in the supporting information) to derive water flow networks, sinks, and drainage catchments on the hydraulic potential surface. Our calculated subglacial drainage pathways and sinks were compared with previous water drainage catchment delineations [*Mernild and Hasholt*, 2009; *Bartholomew et al.*, 2011; *Fitzpatrick et al.*, 2014; *Smith et al.*, 2015], supraglacial lake locations [*Fitzpatrick et al.*, 2014], supraglacial river locations [*Smith et al.*, 2015], and ice flow velocity [*Joughin et al.*, 2010]. Using our subglacial drainage catchment delineations, we have calculated ice sheet surface runoff (melt minus refreezing in snow and firn) using an updated version of the *van As et al.* [2012] surface energy budget model. The runoff was compared to discharge gauged using different techniques for both the Isortoq River and the Watson River. Discharge data from the Isortoq River were measured by *Smith et al.* [2015] using field-calibrated time-lapse photography of braid plain inundation area and a power law correlation between water surface area and discharge [*Gleason et al.*, 2015]. The discharge of the Watson River was measured using observations of river stage and rating curves based on the method of *Hasholt et al.* [2013]. We investigated changes in subglacial catchments by varying the subglacial water pressure between 50 and 110% of ice overburden pressure. Further details of the methods employed are given in the supporting information Text S1.

3. Results

Using hydraulic potential we have characterized subglacial catchments, flow networks, and sinks up to 1750 m above sea level (asl), ~ 120 km from the ice margin (Figure 1). Our study area contains two main subglacial catchments that extend up to the ice divide (Table 1, Figure 1, and supporting information Figure S1), which we refer to as the (1) Isunnguata Sermia and (2) Kangerlussuaq catchments. The Isunnguata Sermia catchment drains into the Isortoq River. The Kangerlussuaq catchment consists of three subcatchments: (2a) the Sandflugtsdalen subcatchment, with the Russell and Leverett outlet glaciers; (2b) the Ørkendalen subcatchment, with the Ørkendalen and Isorlersuup outlet glaciers; and (2c) the Point 660 subcatchment. The rivers emerging from these three subcatchments converge into the Watson River. The calculated subglacial water networks show a dendritic pattern controlled largely by basal topography and which flows from the interior of the ice sheet toward the margin (Figure 1 and supporting information Figures S1–S3). We located 523 subglacial hydrological sinks in the study area, with median and maximum areas of 0.30 km^2 and 12.7 km^2, respectively (supporting information Figure S4). The total modeled runoff from 29 May to

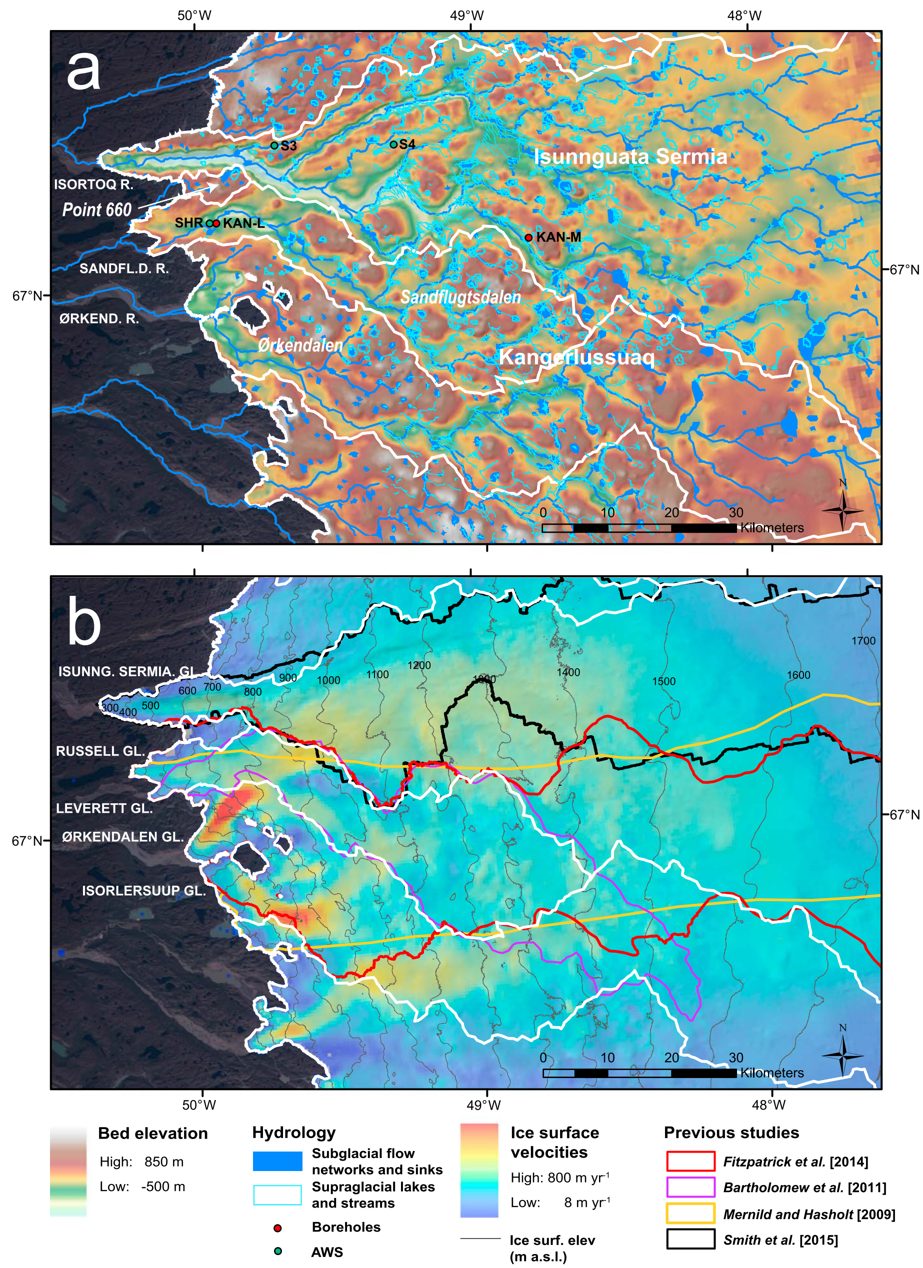

Figure 1. Subglacial hydrology and ice flow. The Isunnguata Sermia and Kangerlussuaq catchments (with three subcatchments: Ørkendalen, Sandflugtsdalen, and Point 660) are delineated with white lines for subglacial water pressures equal to ice overburden. Automatic weather stations [*van As et al.*, 2012] used in the runoff calculations and previous borehole measurement sites [*Meierbachtol et al.*, 2013; *van de Wal et al.*, 2015] are shown. See supporting information Figure S1 for an extended map of the drainage catchments up to the ice divide. (a) Supraglacial lakes (maximum extent 2002 to 2012) [*Fitzpatrick et al.*, 2014] and streams (2012) [*Smith et al.*, 2015] recur in subglacial depressions. See supporting information Figure S2 for a higher-resolution image. (b) Previous delineations of supraglacial [*Mernild and Hasholt*, 2009; *Bartholomew et al.*, 2011; *Fitzpatrick et al.*, 2014] and subglacial (from hydraulic potential) [*Smith et al.*, 2015] catchments correlate spatially with the subglacial drainage catchments close to the ice margin but differ considerably farther inland. Annual ice surface velocities (2005/2006) are from *Joughin et al.* [2010], and surface elevations are from *Howat et al.* [2014]. Elevation is given as height above the WGS-1984 ellipsoid.

10 September 2012 equals $5.4 \pm 0.5\,km^3$ for the Isunnguata Sermia catchment and $6.5 \pm 0.7\,km^3$ for the Kangerlussuaq catchment (Figure 2 and Table S1). Observed total discharge from the Isortoq River was between 2.7 and $5.4\,km^3$ (with linear interpolation used to account for 32 missing observations) [*Smith et al.*, 2015]. Observed total discharge from the Watson River was $6.4 \pm 1.0\,km^3$ from 29 May to 10 September 2012, which

Table 1. Comparison of the Area of Our Subglacial Catchments With Those Defined by Previous Studies Assuming Subglacial Water Pressures Equal to Ice Overburden[a]

	Catchment	Area up to ~ 1750 m asl (km^2)	Area up to Ice Divide ~ 2600 m asl (km^2)
1	Isunnguata Sermia catchment	3,200	15,900
2	Kangerlussuaq catchment	2,800	12,000
	van As et al. [2012]	ND	12,547
	Fitzpatrick et al. [2014]	ND	9,743
2a	Sandflugtsdalen subcatchment	NA	900
	Bartholomew et al. [2011]	NA	1,145
2b	Ørkendalen subcatchment	1,800	11,100
2c	Point 660 subcatchment	NA	40
	Rennermalm et al. [2013a]	NA	31–60

[a]ND, no data; NA, not applicable.

compares well to the modeled runoff for the Kangerlussuaq catchment of 6.5 ± 0.7 km^3. Finally, the modeled runoff from the Sandflugtsdalen subcatchment from 29 May to 29 August 2012 totaled 2.3 ± 0.2 km^3, which compares favorably to the measured discharge of 2.2 km^3 from the outlet river of Leverett Glacier [*Tedstone et al.*, 2013]. Our results (Figure 3 and supporting information Figures S5 and S6) also reveal a significant transience in the delineation of drainage catchments by varying the subglacial water pressure (expressed by factor k in equation (1); see supporting information Text S1), especially within ± 10% of the ice overburden pressures ($k = 0.9$ to 1.1). This range in subglacial water pressure is consistent with borehole measurements in the area; *Meierbachtol et al.* [2013] measured subglacial water pressure factors within the Isunnguata Sermia catchment ranging from $k = 0.88$ to above 1 in the S3 and S4 boreholes, and *van de Wal et al.* [2015] measured water pressures in the Kangerlussuaq catchment ranging from $k \approx 0.95$ to above 1 in the SHR borehole (Figure 1a).

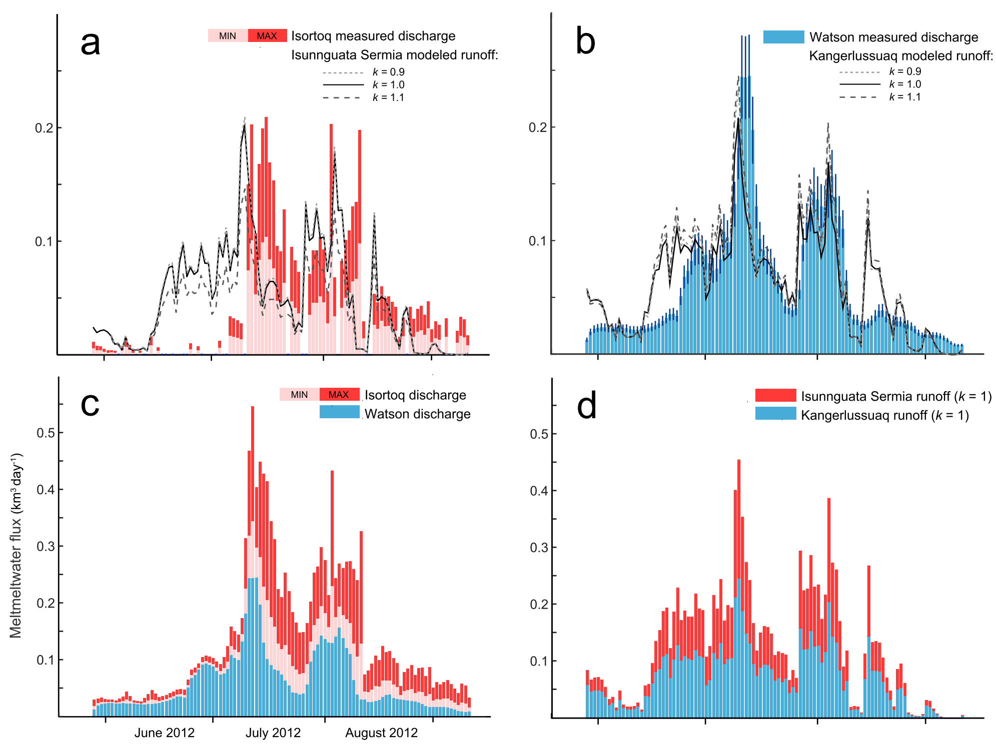

Figure 2. Daily totals of modeled supraglacial runoff and measured proglacial discharge from 29 May to 10 September 2012. (a) The Isunnguata Sermia catchment and the Isortoq River, with maximum and minimum discharge values [*Smith et al.*, 2015]. (b) The Kangerlussuaq catchment and the Watson River, with ± 15% uncertainty in the discharge calculations in error bars. (c) Stacked discharge for the Isortoq (linearly interpolated to fill the 32 missing observations) and Watson proglacial rivers. (d) Stacked runoff for the Kangerlussuaq and Isunnguata Sermia catchment for subglacial water pressures equal to ice overburden.

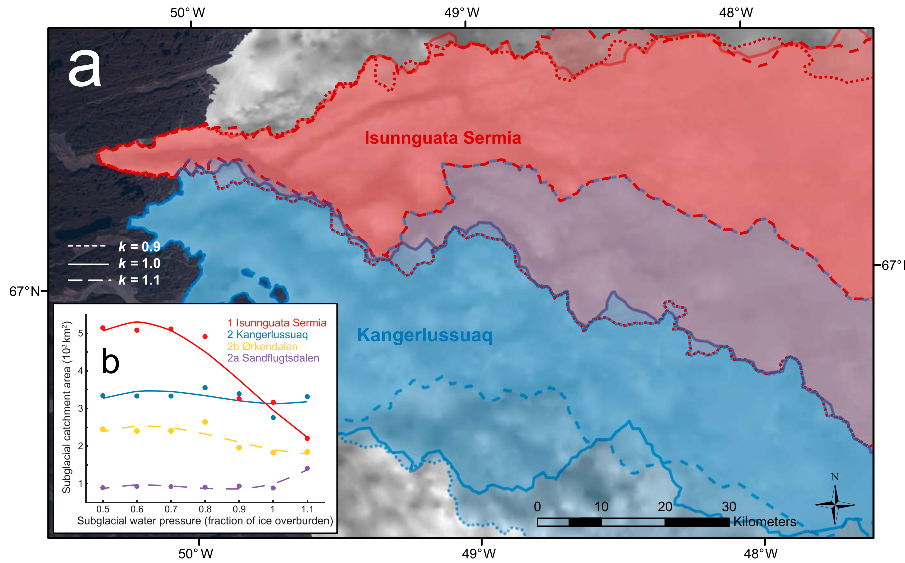

Figure 3. (a) Spatial changes in subglacial drainage delineations at different subglacial water pressures (see equation (1) in the supporting information Text S1) for $k = 0.9$ (dotted line), 1.0 (solid line), and 1.1 (dashed line): the Isunnguata Sermia catchment (red), the Kangerlussuaq catchment (blue), and overlapping areas (purple). See also supporting information Figure S5. (b) Changes in subglacial catchment area with changing subglacial water pressures for $k = 0.5$ to 1.1. The Kangerlussuaq catchment shifts northward for $k = 1.1$ (without changing significantly in size) and takes a large part (~30%) of the area from the Isunnguata Sermia catchment. The area changes are calculated up to 1750 m asl. A cubic polynomial regression was applied to the catchment areas and the coefficient of determination (R^2) was 0.95 for Isunnguata Sermia and 0.38 for Kangerlussuaq (with 0.79 for the Ørkendalen subcatchment and 0.91 for the Sandflugtsdalen subcatchment).

4. Discussion

The hydraulic and ice thickness gradients along the main flow paths in the ablation zone are on the order of -10^1 to -10^2 m km^{-1} (supporting information Figure S3 and Table S2), which is high in comparison with flow paths in Antarctica, where typical values of -10^{-4} to -10^{-3} m km^{-1} have been measured for the hydraulic head and -10^{-3} to -10^{-2} m km^{-1} for the ice thickness [*Wolovick et al.*, 2013]. This can be explained by the relatively rough bed topography [*Lindbäck and Pettersson*, 2015] and steep ice surface slopes in the marginal areas of our study area. Consistent with previous work [*Joughin et al.*, 2013] our results suggest that the spatial pattern of enhanced summer ice flow can be attributed to surface and bed slopes acting in unison to determine subglacial hydrological pathways. Our identified sinks in the hydraulic potential surface (Figure 1a and supporting information Figures S2 and S4) are possible locations for subglacial lakes, and expectedly the density of sinks is much higher than found by previous studies from the GrIS [e.g., *Livingstone et al.*, 2013] that used lower resolution DEMs. Although many of the sinks are small (median area 0.30 km^2), they have a combined area of 290 km^2 and an estimated combined storage capacity of 4.1 km^3. The preponderance of sinks indicates that small subglacial lakes (<1 km^2) are likely to exist in our study area but remain undetected by remote sensing data owing to a negligible surface expression created by subglacial lakes of this size [e.g., *Langley et al.*, 2011]. In many locations the sinks are mirrored at the surface by supraglacial lakes (~40% overlap; Figure 1a), which provides support for the modeling work of *Sergienko* [2013], where supraglacial and subglacial lakes are colocated originating from the same basal features.

Previous studies that delineated hydrological catchments using ice surface topography [e.g., *Mernild and Hasholt*, 2009; *Mernild et al.*, 2010; *Bartholomew et al.*, 2011; *van As et al.*, 2012; *Chandler et al.*, 2013; *Cowton et al.*, 2013; *Fitzpatrick et al.*, 2014] agree well with our subglacial catchments close to the margin (<50 km distance and < 1200 m asl), but large differences occur farther inland (Figure 1b). In accordance with previous studies [e.g., *Budd and Carter*, 1971; *Gudmundsson*, 2003], we suggest that the large differences between previous catchment delineations above 1200 m asl may, in part, be explained by thicker ice having a larger influence on the hydrostatic pressure as well as muting the surface expression of the bed, making accurate delineation of the subglacial drainage from surface topography more difficult. Large discrepancies also occur between our subglacial catchment delineation and the subglacial catchments calculated from the

1 km gridded *Bamber et al.*'s [2013a] DEM [e.g., *Smith et al.*, 2015]. We attribute the differences to our higher data density, which provides greater detail in our ice thickness DEM that is especially important in regions where subglacial water routing is sensitive to small variations in elevation. Moreover, the geostatistical interpolation model that we used to generate the DEM was adapted to the regional setting whereas the *Bamber et al.*'s [2013a] DEM was optimized for the whole GrIS.

The high level of agreement between modeled runoff ($6.5 \pm 0.7\ km^3$) for the Kangerlussuaq catchment and measured bulk proglacial discharge ($6.4 \pm 1.0\ km^3$) in the Watson River suggests that a limited volume of water (~$0.1\ km^3$) is retained after the melt season in this catchment. The coherence between our observed proglacial discharge and modeled runoff compared to previous studies [e.g., *van As et al.*, 2012; *Rennermalm et al.*, 2013a; *Fitzpatrick et al.*, 2014] can be explained by the reduced uncertainty in catchment delineation, especially in the upper part of the ablation zone (>1200 m asl) from where a significant part of the surface runoff originates (65% in 2012) and the majority of supraglacial lakes are located [*Fitzpatrick et al.*, 2014; *Leeson et al.*, 2014]. The discharge estimates of the Isortoq River have larger uncertainties (discharge with lower and upper bounds ranging from 2.7 to $5.4\ km^3$) [*Smith et al.*, 2015], due to a different gauging technique (see section 2). Our modeled runoff ($5.4 \pm 0.5\ km^3$) is close to the upper limit of the measured discharge but exceeds the lower limit by 50%. Previous findings of *Smith et al.* [2015] measured Isortoq discharge to total 37 to 75% of modeled runoff in 2012, suggesting substantial subglacial water storage in the Isunnguata Sermia catchment, as opposed to the evidence of low storage in the Kangerlussuaq catchment. *Rennermalm et al.* [2013a] also found evidence for meltwater retention (up to 54% of the modeled runoff between 2008 and 2012) in the Point 660 subcatchment, which is small (100 times smaller than the Kangerlussuaq catchment) and therefore has larger relative uncertainty in delineation. The speculated missing water may indeed be retained, as hypothesized by these studies, but alternative explanations also exist, such as groundwater losses and subglacial water piracy, the latter of which is discussed below. Water can be stored interannually when water recharges subglacial sinks that have been flushed out during the melt season. Such storage transience is revealed in our hydraulic potential analysis where subglacial sinks decrease in size with higher subglacial water pressures, decreasing with ~ 25% in total size (equating to an area of ~ $70\ km^2$) from $k = 1$ to $k = 1.1$ (supporting information Figure S5). Comparing daily differences between measured discharge and modeled runoff shows that ~ $1.4\ km^3$ of water has a delayed release within the melt season for both catchments. We attribute this to the temporary retention of water in snow, supraglacial lakes, and the subglacial drainage system.

The bed topography in the Kangerlussuaq sector of the GrIS shows highly variable subglacial trough systems, channelizing the ice flow and enhancing ice deformation. The close agreement between ice flow patterns [*Joughin et al.*, 2010; *Palmer et al.*, 2011] and subglacial water networks in our study area indicates that basal water flow has a primary impact on ice dynamics within the troughs. The extent and magnitude of enhanced summer ice flow and measured discharge varies considerably between the adjacent outlets despite similar geometry and climatic control in the area [e.g., *Fitzpatrick et al.*, 2013]. Outlet glaciers within the Kangerlussuaq catchment accelerate with high magnitude at the beginning of the melt season. By contrast, Isunnguata Sermia has relatively low rates of ice flow and proglacial discharge at the melt season onset compared to later in the season (Figure 2). The low discharge values and ice velocities measured for the Isunnguata Sermia catchment cannot be linked to differences in modeled surface melt in the adjacent Kangerlussuaq catchment, and we hypothesize that this is instead caused by water piracy. The Watson River receives a larger fraction (~75%) of the discharge than the Isortoq River before peak runoff on 13 July 2012, whereas after peak runoff the two proglacial rivers each discharge approximately equal amounts (53% and 47%, respectively). Hence, the relatively larger volumes of water draining through the Kangerlussuaq catchment during melt season onset may explain the pronounced early-season ice flow response of the outlet glaciers of this catchment compared to the Isunnguata Sermia catchment. Instead, Isunnguata Sermia accelerates later in the melt season possibly owing to an expanding catchment area (water piracy) driving increased subglacial water routing and discharge. We test the hypothesis of water piracy by varying subglacial water pressures in our subglacial catchment delineation and modeled runoff (expressed by factor k; Figure 3 and supporting information Figures S5 and S6). Fluctuations in water pressure are common in the study area during the melt season both on temporal and spatial scales [e.g., *Meierbachtol et al.*, 2013; *Andrews et al.*, 2014], and the assumption of uniform subglacial water pressure is not accurate over the whole region or over time. At subglacial water pressures below 80% of the ice overburden pressures

($k = 0.5$ to 0.8), which is expected close (<50 km) to the ice margin [e.g., *Meierbachtol et al.*, 2013], the catchments do not change significantly in size (Figure 3b). The largest changes are for k values from 0.9 to 1.1 and occur where the subglacial valleys are shallow, or where subglacial valleys converge (supporting information Figure S5). Remarkably, the Kangerlussuaq catchment migrates northward for $k = 1.1$ but remains similar in area and captures ~ 30% of the area from the Isunnguata Sermia catchment (Figure 3 and supporting information Figure S5). As the net area of the Kangerlussuaq catchment does not significantly change, the total discharge into Watson River neither increased drastically. On the other hand, Isunnguata Sermia catchment does not migrate northward suffers from reduced area and hence reduced discharge. Catchment migration (or water piracy) caused by changes in subglacial water pressure (k values) over, for example, a summer melt season could therefore provide alternative hypothesis to that of meltwater retention. At this time regional basal water pressures would be high and observations of surface uplift indicate that transient water pressures in excess of overburden do occur [*Shepherd et al.*, 2009]. Individual drainage catchments may therefore fluctuate significantly in size in response to relatively small changes in basal water pressure. Hence, the subglacial routing of water may be influenced not only by the basal topography but also by varying subglacial water pressure; both play a key role in governing subglacial water routing and therefore the ice flow regime of individual outlet glaciers. Our findings suggest that a more sophisticated subglacial hydrology modeling to predict variable subglacial water pressure [e.g., *Werder et al.*, 2013] is warranted especially in sensitive areas in the subglacial topography (i.e., areas that are potential switching points) to understand the nonlinear and transient relationships between subglacial water flow (i.e., its routing and discharge) and ice dynamics over diurnal, seasonal, and annual time scales.

5. Conclusions

The presence of water beneath ice sheets has a fundamental impact on ice flow due to its role as a lubricant either between the ice and its base or between grains of subglacial sediment, and hence, water delivery to the bed plays a key role on the rate of dynamic mass loss to global sea level. We present a high-resolution subglacial hydrological analysis of the land-terminating Kangerlussuaq sector of the GrIS, characterizing subglacial catchments, flow networks, and hydrological sinks. The agreement between observed proglacial discharge and modeled melt runoff exceeds that of previous studies using similar runoff calculations, due to reduced uncertainty in our catchment delineation. Nevertheless, our analysis indicates that many small subglacial lakes may occur in this region and yet would remain undetectable by standard methods using coarse-resolution DEMs, ice surface elevation change, or radar techniques. Furthermore, our findings indicate that drainage catchments may vary not only on glacial time scales of 10^1 to 10^3 years, as determined for ice streams in Antarctica, but also on subseasonal time scales in response to relatively small perturbations in the subglacial water pressure. These findings should be considered in studies that attempt to relate estimates of surface runoff from energy balance models with measurements of proglacial discharge and ice dynamics. Our study highlights the need not only for accurate high-resolution DEMs but also for regionally optimized interpolation when conducting detailed hydrological studies of the GrIS. Continued efforts targeting the hydrological system of the ice sheet should over time result in finer spatial coverage, allowing for a broader understanding of the ice sheet's hydrological and dynamic response to climate change and ultimately its contribution to global sea level rise.

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