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Understanding Greenland ice sheet hydrology using an integrated multi-scale approach

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Abstract

Improved understanding of Greenland ice sheet hydrology is critically important for assessing its impact on current and future ice sheet dynamics and global sea level rise. This has motivated the collection and integration of \textit{in situ} observations, model development, and remote sensing efforts to quantify meltwater production, as well as its phase changes, transport, and export. Particularly urgent is a better understanding of albedo feedbacks leading to enhanced surface melt, potential positive feedbacks between ice sheet hydrology and dynamics, and meltwater retention in firn. These processes are not isolated, but must be understood as part of a continuum of processes within an integrated system. This letter describes a systems approach to the study of Greenland ice sheet hydrology, emphasizing component interconnections and feedbacks, and highlighting research and observational needs.

Keywords: ice sheet, hydrology, Greenland

1. The need for treating ice sheet hydrology as an integrated system

The Greenland ice sheet is the largest body of permanent ice and snow cover in the Northern Hemisphere spanning over 1.7 million square kilometers (Bamber \textit{et al} 2001), and is sensitive to changes in regional and global climate. Between 1958 and 2010, Greenland ice sheet mass losses are estimated to have almost tripled, increasing from 110 ± 70 Gta\textsuperscript{−1} (in 1958, Rignot \textit{et al} 2008) to 263 ± 30 Gta\textsuperscript{−1} (between 2005 and 2010, Shepherd \textit{et al} 2012). Since 1996, roughly 50–61\% of these losses are attributed to ice discharge, with the remainder explained by negative surface mass balance from a combination of precipitation variability, runoff losses, and sublimation fluxes (Rignot \textit{et al} 2008, Van den Broeke \textit{et al} 2009). All these processes are strongly connected to...
Greenland ice sheet hydrology and have implications for future sea level rise. Should recent (1992–2009) total mass loss rates of $+21.9 \pm 1$ Gt a$^{-1}$ from Greenland ice sheet alone continue, global sea level will rise $+9 \pm 2$ cm by 2050 (Rignot et al 2011).

The Greenland ice sheet hydrologic system is complex. Each summer, it becomes activated as meltwater is produced on its surface, evaporates and/or ice sublimes into the atmosphere, percolates into firm layers, and feeds runoff into supraglacial lakes, streams and rivers to the ice sheet’s margins (figure 1). Water penetrating firm layers is retained if it refreezes (Pfeffer et al 1991, Reeh 1991, Greuell and Konzelmann 1994, Boggild et al 2005, Boggild 2007), or stored temporarily until critical saturation for runoff formation is achieved (Pfeffer et al 1991). A substantial fraction of surface runoff in the ablation zone may drain into crevasses where it can reach deeper into the ice sheet interior, perhaps even reaching the bed (McGrath et al 2011). The remaining runoff flows through supraglacial stream networks into moulins for rapid vertical transport deeper into the ice (McGrath et al 2011), or temporarily forms lakes on the ice sheet surface (Echelmeyer et al 1991). Supraglacial lakes mostly occur on the western part of the Greenland ice sheet (Sundal et al 2009), where some lakes can empty rapidly and refill thus episodically propagating meltwater to the bed (Zwally et al 2002, Das et al 2008) through the melt season. While it is clear that surface meltwater entering moulins, crevasses, and fractures is routed to the bed through a combination of storage elements and transport pathways, both englacial and subglacial drainage systems are imprecisely known. At the glacier margin, meltwater emerges in proglacial rivers and ice-marginal lakes and discharges into fjords along the entire perimeter of the Greenland ice sheet. During non-melt periods in the winter months, any residual meltwater generally refreezes and marginal discharge is greatly reduced. At this time, winter snow accumulation and redistribution becomes the dominant hydrologic process, but sublimation of blowing snow ( Lenaerts et al 2012) as well as rainfall and occasional melting events (Rennermalm et al 2012a) also take place.

Our current limited understanding of Greenland ice sheet hydrological processes is intertwined with unresolved questions of how present and future climate change may influence ice sheet dynamics and potentially trigger positive hydrology–ice dynamics feedbacks. Dynamic losses are known to occur when greater ice velocities increase calving, retreat, and thinning of ice sheet outlet glaciers. These processes also intensify overall melting rates and volume as more ice flows toward lower elevations, where temperatures are warmer and ablation rates are higher (Parizek and Alley 2004). While the most dramatic ice velocity increases are warmer and ablation rates are higher (Parizek and Alley 2008), the most dramatic ice velocity increases are particularly prominent for land-terminating glaciers (Joughin 2008, Shepherd et al 2009, Bartholomew et al 2011a, Sundal et al 2011). In order for surface melting to influence ice sheet velocity, meltwater must reach the bed and raise basal water pressure and increase ice-sliding velocities. Theoretical studies show that rapid fracture propagation to the bed is entirely possible (Van der Veen 2007), and this is supported by observational evidence of marked velocity increases in response to supraglacial lake drainage (Zwally et al 2002, Das et al 2008) as well as pronounced seasonal velocity variability in moulin-dense regions (Van de Wal et al 2008). However, the relationship between velocity and surface melt is not simply linear. In contrast to ice sheet velocities over shorter time periods, annual velocities appear unrelated to ablation rate (Van de Wal et al 2008), which suggests the existence of self-regulating mechanisms tied to subglacial drainage network development (Pimentel and Flowers 2010, Schoof 2010, Colgan et al 2011, Sundal et al 2011).

A central question related to Greenland hydrology is how much meltwater is truly lost to the ocean and thereby ultimately contributes to rising global sea levels. To understand this contribution, four interconnected aspects of the Greenland hydrologic system must be better understood: (1) surface mass balance, (2) meltwater retention, (3) feedbacks between ice sheet hydrology and dynamics, and (4) runoff losses to the ocean at the ice sheet margin. Notwithstanding advances in surface mass balance modeling of the Greenland ice sheet (Mote 2003, Hanna et al 2005, Box et al 2006, Ettema et al 2009, Mernild et al 2009, Fettweis et al 2012, Fitz Gerald et al 2012), confidence in mass balance flux magnitudes remains low due to the poorly understood hydrologic system. This includes uncertainties about precipitation inputs (Ettema et al 2009, Burgess et al 2010), scarcity of in situ validation, incomplete and

![Greenland Ice Sheet Hydrology](https://example.com/greenland-hydrology.png)

**Figure 1.** Schematic cross section of Greenland ice sheet hydrology from the ice margin to the interior divide for a land-terminating outlet where snow, firm, and ice layers are shown in a cutaway view to illustrate internal structures. Surface mass balance terms are shown as stippled arrows. Englacial and subglacial features are particularly poorly known.
uncalibrated near-surface air temperature data (Shuman et al. 2001), and albedo feedbacks from surface melting (Box et al. 2012). Adding to this uncertainty is limited knowledge about meltwater volumes and fluxes retained in supraglacial lakes and internal glacial storage reservoirs (Sundal et al. 2009, Rennermalm et al. 2012a), and meltwater refreezing in firm (Pfeffer et al. 1991, Forster et al. 2012, Harper et al. 2012, Humphrey et al. 2012). Furthermore, feedback mechanisms between ice sheet hydrology and ice sheet dynamics are being examined with modeling (Phillips et al. 2010, Schoof 2010, Colgan et al. 2011, 2012), but require more observational work. Actual meltwater losses can be monitored in rivers and streams, but such observations are sparse especially considering the large number of such streams active during the melt season (Van de Wal and Russell 1994, Mernild and Hasholt 2009, Rennermalm et al. 2012b), and can only partly be estimated using remotely sensed discharge-fjord sediment area/concentration relationships (Chu et al. 2009, 2012, McGrath et al. 2010).

Ice sheet hydrologic elements and processes have many similarities with glaciers and ice caps (consult these comprehensive reviews: Fountain and Walder 1998, Jansson et al. 2003, Hock 2005, Irvine-Fynn et al. 2011). However, it is unclear if Greenland ice sheet hydrology can be considered simply an upscaled version of glacier hydrology (Irvine-Fynn et al. 2011) or whether scale changes produce new and unique behaviors. Similarly, the Greenland ice sheet might not behave as a downscaled Antarctic ice sheet given Greenland’s much more active hydrological system. Here, we argue that a better understanding of Greenland ice sheet hydrology can be obtained by bridging gaps between small-scale process studies, large-scale modeling, and satellite remote sensing efforts. This requires higher density and availability of high-resolution satellite remote sensing products, airborne observations, field observations, and scaling studies. To demonstrate these points, we briefly review components and processes of Greenland ice sheet hydrology and how they fit together in an integrated system, identify critical components where progress is urgently needed, and discuss why a multi-scale approach is necessary to gain better understanding of this very large ice sheet in a changing climate.

2. Components of the hydrologic system of the Greenland ice sheet

2.1. Surface mass balance

Surface mass balance refers to the net balance over an annual period between accumulation from precipitation and ablation from runoff production, sublimation and evaporation. Besides its relevance for its sea level rise contribution, it has important impacts on surface energy balance (Tedesco and Steiner 2011), ice sheet thermal state (Phillips et al. 2010), and ice-flow dynamic processes (Zwally et al. 2002, Sundal et al. 2011).

2.1.1. Snow accumulation. With nearly all precipitation falling as snow at this time across Greenland, snow accumulation is the primary input to the ice sheet surface mass (Box et al. 2006). The general spatial pattern is well established with maximum snow accumulation along parts of the southeast coast and local maxima in northeast and/or northwest corners (Cogley 2004, Hines and Bromwich 2008, Bales et al. 2009, Burgess et al. 2010, Linling et al. 2011). Interannual and high-spatial-resolution accumulation products, however, have low confidence due to scarce availability of in situ validation from ice cores, snow pits (Ohmura and Reeh 1991, Mosley-Thompson et al. 2001, Hawley et al. 2008), and snow depth observations at automatic weather stations (Bales et al. 2009). Remotely sensed snow accumulation may eventually alleviate the need for in situ observations, but is currently experimental (e.g. Drinkwater et al. 2001). Shortcomings in snow accumulation estimates propagate to a rather large uncertainty into ice sheet surface mass balance estimates (see section 2.1.2). Additionally, modeling studies show locally important drifting snow processes in southwest Greenland (sublimation) and southeast Greenland (erosion/redistribution) acting to reduce annual accumulation totals, but are currently unconfirmed by observational studies (Lenaerts et al. 2012).

About 80% of current accumulation variance can be explained by large-scale atmospheric circulation and its interaction with the geometry of the ice sheet (Van der Veen et al. 2001), and is associated with onshore and upslope flow of moist air strongly correlated with the North Atlantic Oscillation (NAO) (Rogers 2004). Global climate models predict that future circulation changes will enhance the meridional flux over the region, which will increase heat and moisture transport to Greenland (Franco et al. 2011). This may lead to heavier snowfall and a positive net mass balance on the eastern side of the ice sheet (Kiilsholm 2003), even if the total net effect for all of Greenland is negative (Fettweis et al. 2012). Additional heat and moisture may stem from enhanced ocean–atmosphere fluxes triggered by reduced sea ice extent and thickness (Serreze et al. 2009, Screen and Simmonds 2010) as the Arctic Ocean transitions to largely ice free during summer months in the 21st century (Stroeve et al. 2012). Already under present climate, sea ice variability covaries strongly with ice sheet surface melting in some regions (Rennermalm et al. 2009). While linkages between sea ice and Greenland ice sheet surface accumulation and ablation are under explored, modeling studies show that sea ice reductions may influence surface climate many 100 km away from the coast (Lawrence et al. 2008, Higgins and Cassano 2009, 2011).

supplied the surrounding oceans with $\sim220$–$550 \, (\pm 86) \, \text{Gta}^{-1}$ ice discharge (assumed equal to surface mass balance) from calving marine-terminating glaciers, and $\sim250$–$264 \, (\pm 26$–$45) \, \text{Gta}^{-1}$ from ice sheet runoff along the ice sheet margin (Hanna et al 2005, Fettweis 2007, Ettema et al 2009, Van den Broeke et al 2009, Mernild and Liston 2012). Since the mid-1990s, velocity has fluctuated significantly for individual outlet glacier (Rignot and Kanagaratnam 2006, Rignot et al 2008, Moon et al 2012), but increased overall (Moon et al 2012) resulting in ice discharge between $\sim440$ and $600 \, \text{Gta}^{-1}$ (Rignot et al 2008, Van den Broeke et al 2009), while ice sheet runoff has steadily increased to $\sim400$–$429 \, (\pm 57) \, \text{Gta}^{-1}$ (Fettweis 2007, Ettema et al 2009, Mernild and Liston 2012) during 2000–10. Runoff mass losses would have been 100% higher since 1996 if not they had not been offset by increased snowfall and refreezing trends (Van den Broeke et al 2009).

The surface energy balance is the main driver for surface mass losses, and is the sum of incoming and outgoing radiative (shortwave and longwave) and turbulent (sensible and latent heat) fluxes modulated by clouds, winds, topography, and other factors (Cuffey and Paterson 2010, Ettema et al 2010). Albedo is particularly important because it controls absorption of incoming solar radiation, which is the main surface energy balance component (Van den Broeke et al 2008, Tedesco et al 2011, Box et al 2012). Furthermore, model sensitivity studies show that albedo parameterization is the most important parameter controlling future surface mass balance estimates (Fitzgerald et al 2012). In concert with low albedo and high potential for energy release from meltwater refreezing in firm, most melt energy is concentrated near the ice margin, but substantial melt energy can be also found far inland, particularly in southwest Greenland (Ettema et al 2010). In the southwest Greenland ablation zone, sensible heat fluxes provides nearly half the melt energy at lower elevations, while net shortwave radiation is the primary driver at higher elevations (Van den Broeke et al 2008, 2011) and in most of the ice sheet ablation zone (Ettema et al 2010). Between 2000 and 2011, average ice sheet melting season albedo displayed a strong decline indicating its growing importance in driving melt anomalies (Box et al 2012). In a future, warmer climate, currently snow covered parts of the percolation zone are likely to expose an underlying darker ice surface resulting in a positive albedo feedback (Box et al 2012, Franco et al 2012) augmented by enhanced sensible heat advection projected to increase positive melt anomalies especially near the ice sheet margin (Franco et al 2012).

Considerable effort needs to be placed on constraining error estimates on reconstructions of past and future surface mass balance components. Surface mass balance estimates are typically made with models of different sophistication levels ranging from positive degree-day models (e.g. Braithwaite 1995), to energy balance models (e.g. Van den Broeke et al 2008). Positive degree-day models relating cumulative positive temperatures to surface melting (Braithwaite 1995, Hock 2003) have been useful in providing first order estimates of present and future Greenland ice sheet surface melting (Mote 2003, Box et al 2004, 2006, Hanna et al 2008). However, energy balance models simulate the actual surface mass balance drivers (Van de Wal and Oerlemans 1994, Van den Broeke et al 2008, Van As et al 2012), which allows for more detailed investigation of ice sheet hydrology, e.g. dependence of firm refreezing on albedo parameterization (Van Angelen et al 2012), blowing snow (Lenaerts et al 2012), and underlying causes for melt extremes (Tedesco et al 2008, 2011).

Model improvements can be made by exploring causes for model disagreement, including initial and boundary conditions, and differences in processes resolved by various models (Fettweis et al 2011, Rae et al 2012), such as retention in cold snow (Pfeffer et al 1991, Greuell and Konzelmann 1994, Janssens and Huybrechts 2000, Boggild et al 2005), ice mask extent (Vernon et al 2012), parameterization of albedo and topography (Rae et al 2012), englacial storage, and snow compaction (Wake et al 2009). Increased process understanding of meltwater storage in firm, lakes, englacial and subglacial storages and hydrological drainage networks (see sections 2.2–2.4) will allow time evolution of spatially distributed runoff to be fully integrated into hydrologic runoff-routing models, and water exchange between surface- and en-, and subglacial environments (see section 2.4). A central challenge for model improvements is sparse data availability for ground truth validation from both automatic weather stations (Steffen and Box 2001, Van de Wal et al 2005, van As and Fausto 2011) and ablation stakes (Andersen et al 2010, Van de Wal et al 2012). This is particularly problematic regarding meltwater percolation, refreezing and storage in firm layers (see section 2.2), and point validation of surface energy and mass balance components (see section 4), while spatially distributed high-resolution albedo and surface and basal topography products are becoming increasingly available (see section 4).

Scarcity of in situ validation data can be partially overcome by extending the surface mass balance record (e.g. Wake et al 2012) to re-evaluate the relationship between surface mass balance and ice dynamics over longer time scales (Hanna et al 2011), using remotely sensed surface melt area extent as independent model validation (Fettweis et al 2011, Mernild et al 2011b) as well as reconciling modeled mass balance losses with temperature anomalies (Hanna et al 2011), and total mass loss determined with satellite observations from the Gravity Recovery and Climate Experiment (GRACE) (Velicogna and Wahr 2005, Ramillien et al 2006, Wouters et al 2008, Van den Broeke et al 2009, Chen et al 2011, Siemes et al 2013). Remotely sensed melt area extent with passive microwave sensors and mass loss determined with GRACE have relatively coarse resolution (25 km and larger) and are thus sufficient for large-scale validation, but offer little insight into surface mass balance processes. Another strategy includes implementing schemes that allow validation with remotely sensed fields, e.g. comparison of albedo determined with remote sensing and modeled as a function of grain size provides insight into grain size model components (Van Angelen et al 2012). Perhaps the greatest potential of currently available remotely sensed data products lies in improving initial and boundary
forcing for surface mass balance models. This includes fields of ice albedo (Lenaerts et al 2012), and could be extended to include radiation balance components from the Clouds and the Earth’s Radiant Energy System (CERES) sensor on NASA’s Aqua and Terra missions (Wielicki et al 1996, Zhou et al 2007), near-surface atmospheric moisture and temperature profiles from the AIRS experiment on NASA’s Aqua mission (Susskind et al 2003), cloud cover with Moderate-Resolution Imaging Spectroradiometer (MODIS), and cloud properties with CloudSat and the Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) (Liu et al 2012).

2.2. Surface meltwater retention and refreezing in firn

In the lower reaches of the accumulation zone, referred to as the percolation zone, surface meltwater is produced and can run off or be retained within the snowpack and/or firn such that runoff leaving the ice sheet is less than the total meltwater production. Meltwater retention and refreezing may be quite substantial. Two widely applied surface mass balance models (MAR and RACMO2) estimate annual average refreezing to 202 and 295 Gt a\(^{-1}\) between 1958 and 2007, which is 45–49% of the total annual melt (see supplemental documentation in Ettema et al 2009). For refreezing to occur, meltwater must percolate deeper into available firn pore space. A first estimate puts percolation zone firm meltwater storage capacity between 322 (±44) to 1289 (−252, +388) Gt (Harper et al 2012). Adding to these numbers is the recent discovery of widespread perennial water saturated firn aquifers (Forster et al 2012). Most runoff is produced in the ablation zone, without significant firn storage capacity, and annual snow fall will replenish firm mass, such that it will take decades to exhaust firm storage capacity (Harper et al 2012). A string of recent years with record widespread ice sheet surface melting in 2005 (Hanna et al 2008), 2007 (Mote 2007, Tedesco 2007a), 2010 (Box et al 2010), and 2012 (Nghiem et al 2012) suggest that buffering of meltwater losses by firn storage may grow in importance as surface melting becomes more frequent in the percolation zone. Thus, high priority should be given to further constrain storage capacity, meltwater percolation, and refreezing in firm.

Modeling has the potential to become an important tool in constraining firm storage capacity and meltwater refreezing in firm. However, current surface mass balance model estimates of firn meltwater refreezing of total melt vary widely from 13–23% in simpler models to 45–49% in more comprehensive models (see supplemental documentation in Ettema et al 2009). To date, relatively little work has been done in determining a robust refreezing scheme for Greenland, partly due to lack of \textit{in situ} observational data for validation (Bougamont et al 2007, Reijmer et al 2012). While sensitivity studies show that refreezing parameter uncertainty in a surface mass balance model are smaller than other parameters (i.e., albedo), it still has regional importance (Fitzgerald et al 2012).

At the same time, more studies are needed to understand the significance of meltwater percolation and refreezing processes.

A key outstanding question is the fate of meltwater as it percolates into firn layers: will meltwater be retained as frozen ice lenses and saturated slush layers, or will it runoff along perched ice lenses and the firm/ice interface? Observational evidence suggests that most meltwater refreezes at depth at higher elevations, while at lower elevations water is more likely to migrate horizontally and escape as runoff (Harper et al 2012, Humphrey et al 2012). In between, a continuum of processes from complete refreezing and runoff takes place, including: (1) inhomogeneous flow or ‘piping’; (2) advancing meltwater front; (3) freezing of meltwater deep into firn (Humphrey et al 2012). Meltwater retained in the upper snow and/or firm layers throughout the summer refreeze during winter, forming a high density ice layer that often can be observed in the upper few meters (Braithwaite et al 1994, Benson 1962, Humphrey et al 2012), and releasing latent energy raising firm temperatures (Pfeffer et al 1991, Reeh 1991, Humphrey et al 2012). At depths (>10 m), year round presence of liquid water has been inferred from temperature observations (Humphrey et al 2012) and confirmed in the field with the recent discovery of extensive perennial firm aquifers (PFA) (Forster et al 2012).

The PFA represents a new storage mechanism for the ice sheet, and needs to be considered in ice sheet mass and energy budget calculations and possibly dynamic models as well. Direct evidence of ice sheet liquid water retention in PFA over winter was discovered in April 2011 in southeast Greenland while drilling two ice cores at ~1600 m elevation (Forster et al 2012). The top of the April 2011 PFA was traced by ground penetrating radar to lie between 8 and 25 m (Forster et al 2012). Airborne radar data from NASA’s Operation IceBridge (OIB) (Koenig et al 2010) acquired 11 days prior to the ground observations and high-resolution coupled regional atmosphere–firm model (Van Angelen et al 2012) showed that PFAs are concentrated in the south and retained 17 ± 1 Gt of water in April 2011 (Forster et al 2012).

2.3. Supraglacial lakes

Supraglacial lakes form during the summer in topographic depressions on the Greenland ice sheet surface, and incise into the surface by roughly doubling ablation rates through lowered surface albedo (Luthje et al 2006, Tedesco et al 2012). These lakes cover large areas, particularly in southwest Greenland where an average of 61% of the total ice sheet became lake covered between 2005 and 2009 (Selmes et al 2011). However, assessments of meltwater volume and fluxes from these lakes are in their infancy, such that the importance of these lakes in terms of net storage and ice sheet dynamic feedbacks (see section 3) is still unknown.

Lake basins may persist from year-to-year or drain episodically (Selmes et al 2011, Liang et al 2012), and sometimes supply meltwater deep into the ice sheet (Das et al 2008). Over the course of the melting season, supraglacial lakes are established in longitudinal bands at increasingly higher elevations (Sundal et al 2009, Lamkin 2011). During warmer years, supraglacial lakes drain more frequently and earlier in the melt season, with the lake population extending...
to higher elevations over the course of the summer (Liang et al. 2012). This suggests that recent increases in surface meltwater production (Tedesco et al. 2008, 2011) may increase the spatial extent, temporal frequency and season of lake drainage events. Assuming this water reaches the bed, this could also influence the spatial pattern of ice sheet velocity (see section 3).

Estimates of lake area and volume can be made with optical satellite imagery (Box and Ski 2007, Sneed and Hamilton 2007), but have only been validated in a relatively small number of studies (Sneed and Hamilton 2007, Tedesco and Steiner 2011). However, there is considerable uncertainty associated with remotely sensed estimates of lake volume because of the assumed impact of optical parameters (e.g., bottom albedo, water attenuation factor) on lake depth retrieval (Box and Ski 2007). One way to improve estimates of the lakes volume and their role in the hydrologic system is through the combination of in situ measurements of bathymetry, satellite data and modeling (Banwell et al. 2012).

2.4. Ice sheet hydrologic network

The Greenland ice sheet hydrologic network is likely to play an important role in regulating feedbacks between hydrology and ice sheet dynamics (see section 3), but also dictates where meltwater and sediment exit the ice sheet to surrounding oceans. Exploration of these features are just beginning, and aside from modeling includes mapping of supraglacial channels using high-resolution remote sensing observations (Yang and Smith 2013) and identification of moulins and englacial structures with ice penetrating ground radar (Catania et al. 2008, Catania and Neumann 2010).

Modeling hydrological networks allows investigations of connections between meltwater production, input into the subglacial environment, and ice dynamics, but also helps constrain true ice sheet meltwater flux to oceans. Two distinct model types are being developed to better understand the importance of the ice sheet’s hydrological connections. The first type of model represents linkages between supra and subglacial hydrology (Flowers and Clarke 2002), which are essential to understand interactions between surface mass balance and ice sheet dynamics (see section 3). Currently, this type of model has only been applied to small parts of the ice sheet (Colgan et al. 2011) or in idealized experiments (Pimentel and Flowers 2010, Schoof 2010, Hewitt 2011). These first model implementations show how meltwater production and propagation to the bed and seasonal development of the subglacial drainage system are linked to ice sheet velocity speedups, an important unknown in sea level rise projections (see section 3).

The second type of model characterizes large-scale drainage patterns for the entire ice sheet using ice sheet topography datasets, either most simply from the shape of the hydrostatic pressure field by also incorporating basal bedrock topography, (Lewis and Smith 2011), or surface topography combined with a network of linear reservoirs (Liston and Mernild 2012, Mernild and Liston 2012). Linear reservoir modeling can be traced back to classical hydrologic theory (Brutsaert 2005) and has been applied at smaller scales to model meltwater runoff from glaciers (Hock and Noetzli 1997, Klok et al. 2001, Verbunt et al. 2003, Hock et al. 2005) and the Greenland ice sheet (Van de Wal and Russell 1994). The primary benefit of these models is perhaps their capability to examine interannual and melting season network dynamics (Lewis and Smith 2011), and to estimate where and when meltwater exits the ice sheet (Liston and Mernild 2012, Mernild and Liston 2012). However, greater importance of moulins relative to crevasses to transmit meltwater (McGrath et al. 2011), and strong correlation between late summer ice sheet velocity change and supraglacial channelization patterns (Palmer et al. 2011) suggest that mapping and modeling drainage network structures at high resolution may be central to understand feedbacks between ice sheet hydrology and dynamics.

2.5. Understanding ice sheet hydrology through analysis of river discharge

River discharge provides direct observations of how much meltwater actually escapes the ice sheet, and has great potential for calibrating and validating surface mass balance models (Mernild et al. 2011a, Van As et al. 2012). This can provide information regarding seasonal development of the subglacial drainage system (Bartholomew et al. 2011b, Bhatia et al. 2011), capturing catastrophic drainage events (referred to as jökulhlaups) that are a common occurrence in the Greenland proglacial zone (Mernild and Hasholt 2009, Mernild et al. 2008, Russell et al. 2011, Russell 1989, 2009), and assessment of meltwater retention (Rennermalm et al. 2012a). Unfortunately, very few rivers have been and are currently being monitored in Greenland especially as a percentage of the total number (Van de Wal and Russell 1994, Mernild and Hasholt 2009, Rennermalm et al. 2012b) because of remoteness as well as due to difficulties with measuring discharge in braided, and highly turbid rivers.

Further insight into glacial hydrologic drainage systems is possible using tracers such as radiogenic, cosmogenic isotopes, and stable isotopes of water, naturally occurring elements, and meltwater physicochemical properties. These approaches have long been used for alpine glaciers (Tranter and Raiswell 1991, Theakstone and Knudsen 1996, Tranter et al. 1996, e.g. Fountain and Walder 1998, Lyons et al. 2002), and are increasingly being applied to Greenland. For example, variations in electrical conductivity can identify links between subglacial lake drainage and peaks in proglacial river discharge (Bartholomew et al. 2011b). Stable isotope composition (δ18O, δD) of meltwater runoff may also be useful, because isotopic compositions in snow reflect atmospheric circulation patterns and temperatures, such that isotope composition correlates strongly with elevation and distance from the ocean (Dansgaard 1964, Dansgaard et al. 1993, Vinther et al. 2010), and can be used to constrain drainage areas by the ice margin (Reeh and Thompson 1993).

Flow paths may be traced using conservative tracers such as Na+, Cl−, Ca2+, and Sr2+, which are highly soluble and have low affinity, thus allowing their concentration to be used as an indication of channelized versus distributed drainage systems.
(Tranter et al. 2002), and meltwater storage in non-melting
winter months (Wadhams 2000). Meltwater source areas
and ages may be established with non-conservative isotope
tracers such as \(^{210}\)Pb, and \(^{7}\)Be, which have lower solubility
and high tendency to adhere to particle surfaces or form
coils. Strontium isotope ratios \(\left({^{87}\text{Sr}/^{86}\text{Sr}}\right)\) are stable,
unique to rocks and minerals, and can identify weathering
reactions (Lyons et al. 2002). \(\left({^{87}\text{Sr}/^{86}\text{Sr}}\right)\) has successfully
been used to identify source areas of solutes and sediments
for a proglacial stream in Greenland by tracing subglacial
erosion areas or differentiating between soluble input from
atmospheric dust versus bedrock (Hagedorn and Hasholt
2004). Finally, isotopes such as \(^{7}\)Be, and \(^{210}\)Pb provide
insights into age of water and dust, and of residence time
of surface meltwater within the glacial system. Seasonal
variations in the cosmogenic isotope \(^{7}\)Be can quantify
contribution of fast flow-dominated surface meltwater, while
\(^{222}\)Rn, a uranium decay product in rocks and sediments, may
reveal changes from a cavity-dominated to channel-dominated
subglacial drainage systems (Bhatia et al. 2011, Kies et al.
2011). \(^{22}\)Na is another cosmogenic isotope that, if current
limitations are overcome, could become an important method
in understanding meltwater flow pathways (currently, analysis
of \(^{22}\)Na requires collection of an unfeasible, >500 l, amount
of water as well as its transport to a laboratory) (Zhang
et al. 2010).

2.6. Understanding of Greenland ice sheet hydrology with
remote sensing techniques

The scarcity of data of direct mass losses such as river
discharge data requires other indicators of meltwater runoff
to be explored, such as remote sensing techniques. Fjord
sediment plumes, visible in remotely sensed images, provide
a link between meltwater produced on the ice sheet surface
and meltwater released to the ocean (Chu et al. 2009,
McGrath et al. 2010). Sediments produced as ice move over
bedrock are flushed out with meltwater to form buoyant
freshwater plumes floating over more dense saline marine
water. These plumes are most readily detected downstream
of rivers draining land-terminating glaciers, owing to high
suspended sediment concentration and minimal obstruction
by calving ice. For such rivers, excellent agreement between
plume length and discharge demonstrate how this method
can capture meltwater variability averaged over a few
days (McGrath et al. 2010). On a larger scale, ice sheet
meltwater release from marine-terminating glaciers can also
be determined. Comprehensive analysis of remotely sensed
sediment fluxes from all of Greenland illustrate highest
meltwater losses in Southwest Greenland (Chu et al. 2012)
consistent with the spatial distribution of surface runoff and
drainage area configuration (Lewis and Smith 2011). The
greatest promise of this method lies perhaps in inferring
meltwater discharge variability, as absolute losses require
site-specific calibration. However, before this method can be
systematically applied, importance of site-specific conditions
must be established. This includes braided river plain
influence on sediment transport, relationship between fjord
geography/bathymetry and salinity, and how glaciological
variables such as glacier size, sliding speed, ice flux, erosional
susceptibility of bedrock, and meltwater production influence
sediment production (Hallet et al. 1996).

Another unexplored avenue for examining meltwater
loss variability is using high-resolution remote sensing
observations to assess braided river width. River width,
similar to river depth and velocity, can typically be expressed
as a power-law function of discharge (Leopold and Maddock
1953). This technique has been applied successfully to Arctic
braided river systems (Ashmore and Sauks 2006, Smith et al.

3. Potential feedbacks and linkages between ice
sheet hydrology and dynamics

Ice sheet hydrology and dynamics are linked in three ways.
First, enhanced ablation along the margin (and perhaps more
accumulation in the interior) (see section 2.1.1) increase the
ice sheet slope, which is the primary driver of deformational
ice sheet movement (Cuffey and Paterson 2010). Second,
enhanced meltwater input into the ice sheet can induce cryo-
hydrologic warming leading to expanding area of thermal bed
conditions (Phillips et al. 2010) which in turn may increase
sliding velocities. Finally, increased meltwater input into the
ice sheet hydrologic system could overwhelm subglacial
transmission capacities resulting in basal sliding and speed
up. All three processes may bring more ice mass to lower
elevations with amplified ablation rates resulting in a positive
feedback effect. However, this positive feedback effect is
brought into question by recent observations and modeling
studies suggesting that development of an efficient subglacial
meltwater transportation system essentially self-regulates the
speedup.

Early models coupling ice sheet dynamics with surface
energy balance demonstrate enhanced losses when basal
sliding is considered (Van de Wal and Oerlemans 1997). Later
modeling studies show that the efficiency of basal sliding is
modulated by two processes: (1) englacial water transmission
rates (McGrath et al. 2011), and (2) the state of the subglacial
hydrologic system (Pimentel and Flowers 2010, Colgan et
al. 2011). Regarding the first process, it is known that ice velocity
may increase five- to ten-fold in response to episodic lake
drainage (Hoffman et al. 2011) for portions of the year,
but can also speed up in response to continuous meltwater
input (Pimentel and Flowers 2010, Colgan et al. 2011).
Modeling studies suggest that fast meltwater propagation
through moulins is more likely to exhaust basal drainage
capacity and result in basal sliding than slow flow seeping
through crevasses and fractures (McGrath et al. 2011).

The ice sheet subglacial hydrologic drainage systems
can be described as bi-stable systems, where one state is
channelized flow that is highly efficient at transporting water,
and the alternate state is slow flow through interconnected
cavities formed at the ice/bedrock interface, and/or incised
bedrock channels with permeable bed sediments (Kamb et al.
1994, Fountain and Waldner 1998, e.g. Cuffey and Paterson
2010, Sundal et al. 2011). According to this theory, early
melting season subglacial water flows mostly through cavities with low transmission capacity. At this time, large meltwater inputs may overwhelm basal water transport capacity and trigger basal sliding as a result of bed separation and reduced basal friction (Iken and Bindschadler 1986, Anderson 2004, Bartholomau et al 2008). As the melting season progresses, an efficient channelized subglacial drainage system evolves that allows greater meltwater throughput. This explains why ice sheet velocities may decelerate despite further surface melting (Schoof 2010). Observations of seasonal shifts from cavity to channel-dominated flow have been inferred from velocity measurements made with global positioning systems (Bartholomew et al 2010), remote sensing (Sundal et al 2011), isotope studies (Bhatia et al 2011), and shown by modeling studies (Pimentel and Flowers 2010, Colgan et al 2011).


4. Toward an integrated multi-scale approach for better understanding of Greenland ice sheet hydrology

A major motivation to study Greenland ice sheet hydrology is its current and likely future impact on global sea levels (Shepherd et al 2012). Meltwater flux from the ablation zone escaping at the ice sheet margin are controlled by surface mass balance, meltwater retention, and how much runoff is channeled through networks of supra-, en-, and subglacial cavities and channels. While some supraglacial channels drain at the ice sheet margin into rivers, fjords and lakes, frequent drainage occurs through moulins and crevasses into englacial and subglacial environments. In englacial and subglacial environments, additional water is supplied from melting of channel walls. Eventually, meltwater routed through the hydrologic network reaches the ice sheet margin and is delivered to lakes, rivers, and fjords, ultimately reaching the ocean. In addition, meltwater runoff also contributes to raising sea levels augmented by ice discharge at calving glaciers, a type of dynamic loss, modulated by oceanic drivers (e.g. Holland et al 2008), and the state of the subglacial system.

Three issues are particularly important to resolve to better constrain future ice sheet hydrologic influence on sea levels. These are: (1) albedo feedbacks; (2) feedbacks between ice sheet hydrology and dynamics; and (3) meltwater retention in firm. Surface albedo plays a fundamental role in both seasonal and long-term ice sheet meltwater export by modulating net solar radiation, the single most important melt driver in the ablation zone (see section 2.1.2). Positive feedback interactions between hydrology and ice dynamics could amplify meltwater losses and are currently not accounted for in model projections (see section 3). Finally, firm meltwater storage has a huge potential to buffer meltwater losses for decades (see section 2.2), and improved confidence in modeling of this process is urgently needed (see section 2.1.2). These three processes’ influence on meltwater export to the ocean must be estimated and modeled before accurate projections about future sea level changes from the Greenland ice sheet can be made. However, because these processes do not operate in isolation, but are interconnected through various hydrologic pathways and processes manifested on different spatial scales and configuration, a multi-scale integrated approach is needed.

Ice sheet surface albedo is modulated by all surface hydrologic processes. Over the melting season, all these factors collectively act to darken the ice sheet surface through refreezing in firn (larger grain sizes result in lower albedo), expansion of supraglacial lakes and stream networks, migration of the snow/firm (high albedo) line to higher elevations, and concentration of dust and sediment (low albedo) on the ice surface as they melt out. The only exceptions to melt season albedo reduction are occasional summer snowfall events that lighten the ice sheet surface and increase albedo.

Linkages between hydrology and ice sheet dynamics appear to be modulated by the state of the subglacial drainage system (see section 3), but also by englacial transport system efficiency (see section 2.4). Regardless of the state of the bi-stable subglacial hydrologic network, ice-marginal areas tend to respond faster to surface melting than thicker inland areas, suggesting amplified future dynamic meltwater losses from land-terminating outlet glaciers (Bartholomew et al 2011a) as melt intensity and duration increases inland across the ice sheet (Hanna et al 2008, Mermild et al 2010a), and melts snow to expose darker ice surfaces (Fettweis et al 2012).

Firm percolation is intimately linked to the runoff production term of the surface mass balance (see section 2.2). Meltwater percolating into the firm may be retained in pore space, form aquifers, or refreeze as ice lenses (see section 2.2). Firm meltwater volume is controlled by firm storage capacity and surface mass balance. However, only part of the total meltwater production ends up in firm layers with the remainder being routed to the supraglacial hydrological network in lakes, crevasses, streams, and moulins (see section 2.4). Thus, understanding of this process is aided by better-constrained surface mass balance flux partitioning, and meltwater export estimates.

To advance our understanding of these components of the hydrological systems all modes of geophysical exploration are needed, including in situ observations, modeling studies, and incorporation of climate reanalysis data, airborne, and remotely sensed datasets. To this end, the cryosphere science community has benefited tremendously from data sharing, for example in situ weather station networks on and off the ice sheet, including well established, ongoing efforts such as Danish Meteorological Institute (e.g. Laursen 2010) and Greenland Climate Network’s automated weather stations (AWS) (Steffen and Box 2001), and ablation stakes along the so called K-transect in west Greenland (Van de Wal et al 2005, e.g. Van den Broeke et al 2008, 2011), the
PROMICE network (van As and Fausto 2011), and coastal AWS operated by Danish Meteorological Institute and Asiaq available at National Climate Data Center (National Climate Data Center 2012). Many of these networks provide real time observations, and are often used in surface mass balance modeling efforts as either independent validation (e.g. Ettema et al 2009) or forcing inputs (e.g. Mernild et al 2010b) (section 2.1.2). In situ studies also provide unique datasets that are becoming increasingly available online, for example proglacial river discharge (e.g. Rennermalm et al 2012b), surface mass balance (Van de Wal et al 2012), and ice cores (Bales et al 2001, 2009). All of these help constrain error estimates of surface mass balance components.

At this time, the exact roles and functions of internal and supraglacial hydrologic pathways and water storage remains a knowledge void and scientific frontier, which limits predictive modeling of ice sheets in a changing climate. However, increasingly sophisticated models are being developed and applied to examine details of the hydrologic system, for example, the en- and subglacial drainage system (see section 2.4). Collaborations around model output sharing and comparison are frequently seen (Fettweis et al 2011, Rae et al 2012, Reijmer et al 2012, Vernon et al 2012), which helps assessments of forcing data quality, algorithm development and improved representation of processes such as firm refreezing and storage.

Growing availability and diversity of airborne and satellite remote sensing data since the early 1970s are accelerating the subfield of ice sheet hydrology toward new discoveries. Spaceborne passive microwave sensors are instrumental for identifying surface melt timing and extent, (Abdalati and Steffen 2001, Mote 2007, Tedesco 2007b) (section 2.1.2). Total ice sheet mass loss is assessed with the GRACE satellite mission since its launch in 2002 (section 2.1.2). Ice sheet surface properties, including albedo (Hall et al 2002, Stroeve et al 2006, Hall and Riggs 2007), surface temperature (Hall et al 2012), and supraglacial lake distribution (Liang et al 2012) are determined using MODIS. High-resolution products from Landsat, Advanced Spaceborne Thermal Emission and Reflection radiometer (ASTER), and WorldView among other sensors are used to examine supraglacial lake processes (Box and Ski 2007, Tedesco and Steiner 2011), and even supraglacial streams (Yang and Smith 2013). In Antarctica, subglacial lake changes have been successfully identified with satellite and airborne laser altimetry (Smith et al 2009, Scambos et al 2011) suggesting that this technology is suitable for Greenland as well. Observations of en- and subglacial properties such as firn aquifers (Forster et al 2012), drainage structures (Catania et al 2008), layering, and subglacial topography are being facilitated by ground-based ground penetrating radar (GPR) and airborne radars developed and operated by Center of Remote Sensing of Ice Sheets (CresSIS) established by the National Science Foundation at the University of Kansas (Gogineni et al 2001, Leuschen et al 2010, Paden et al 2010), and NASA Operation IceBridge (Koenig et al 2010, Studinger et al 2010) accessible at the National Snow and Ice Data Center (NASA Operation IceBridge 2012).

Imperative to understanding any ice sheet’s hydrologic system is integrating local processes to the scale of the entire ice sheet. To this end, better knowledge of how best these processes should be upscaled is needed. Several promising advances have been made in this direction, including assessing the extent and volume of firn aquifers using ground based and satellite radar observations in combination with surface mass balance models (Forster et al 2012) (section 2.2), and mapping supraglacial lakes from multi-temporal satellite data using automated schemes (Sundal et al 2009, Liang et al 2012) (section 2.3). Newer remote sensing technologies providing very high-resolution images with meter-submeter scale resolution will undoubtedly provide insights into scaling behavior of hydrologic processes such as supraglacial streams and lakes (e.g. Tedesco et al 2012, Yang and Smith 2013). One particular upscaling problem is en- and subglacial hydrology due to the inaccessibility of these environments. Scaling and representation of en- and subglacial hydrology will benefit from continued process studies, and advances in modeling of en- and subglacial networks (see section 2.4) in combination with observations of proglacial river discharge with or without tracers (see section 2.5).

5. Conclusions

In this letter, we present a framework for understanding Greenland’s meltwater hydrology as an interconnected system of multi-scaled processes. These processes form a set of complex interactions but may be separated into four key hydrologic components necessary to understand this important ice sheet’s contributions to present and future global sea level rise: i.e. surface mass balance, meltwater retention, positive feedbacks between hydrology and ice dynamics, and ice sheet margin meltwater losses. Greenland ice sheet hydrology is an exciting, rapidly developing subfield at the nexus of cryospheric and hydrologic science, aided by increased availability of geophysical data as well as new and more capable technologies for remotely sensed observations. Key challenges remain in understanding albedo feedbacks, ice sheet hydrology dynamic feedbacks, and meltwater retention in firm, as well as bridging spatial scales and understanding scaling behavior of small-scale processes, and maintaining consistent observational data sets, both in situ and remotely.

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