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Supraglacial Streams and Rivers

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Abstract

Supraglacial meltwater channels that flow on the surfaces of glaciers, ice sheets, and ice shelves connect ice surface climatology with subglacial processes, ice dynamics, and eustatic sea level changes. Their important role in transferring water and heat across and into ice is currently absent from models of surface mass balance and runoff contributions to global sea level rise. Furthermore, relatively little is known about the genesis, evolution, hydrology, hydraulics, and morphology of supraglacial rivers, and a first synthesis and review of published research on these unusual features is lacking. To that end, we review their (*a*) known geographical distribution; (*b*) formation, morphology, and sediment transport processes; (*c*) hydrology and hydraulics; and (*d*) impact on ice sheet surface energy balance, heat exchange, basal conditions, and ice shelf stability. We conclude with a synthesis of key knowledge gaps and provide recommendations for future research.

- Supraglacial streams and rivers transfer water and heat on glaciers, connecting climate with subglacial hydrology, ice sliding, and global sea level.
- Ice surface melting may expand under a warming climate, darkening the ice surface and further increasing melt.

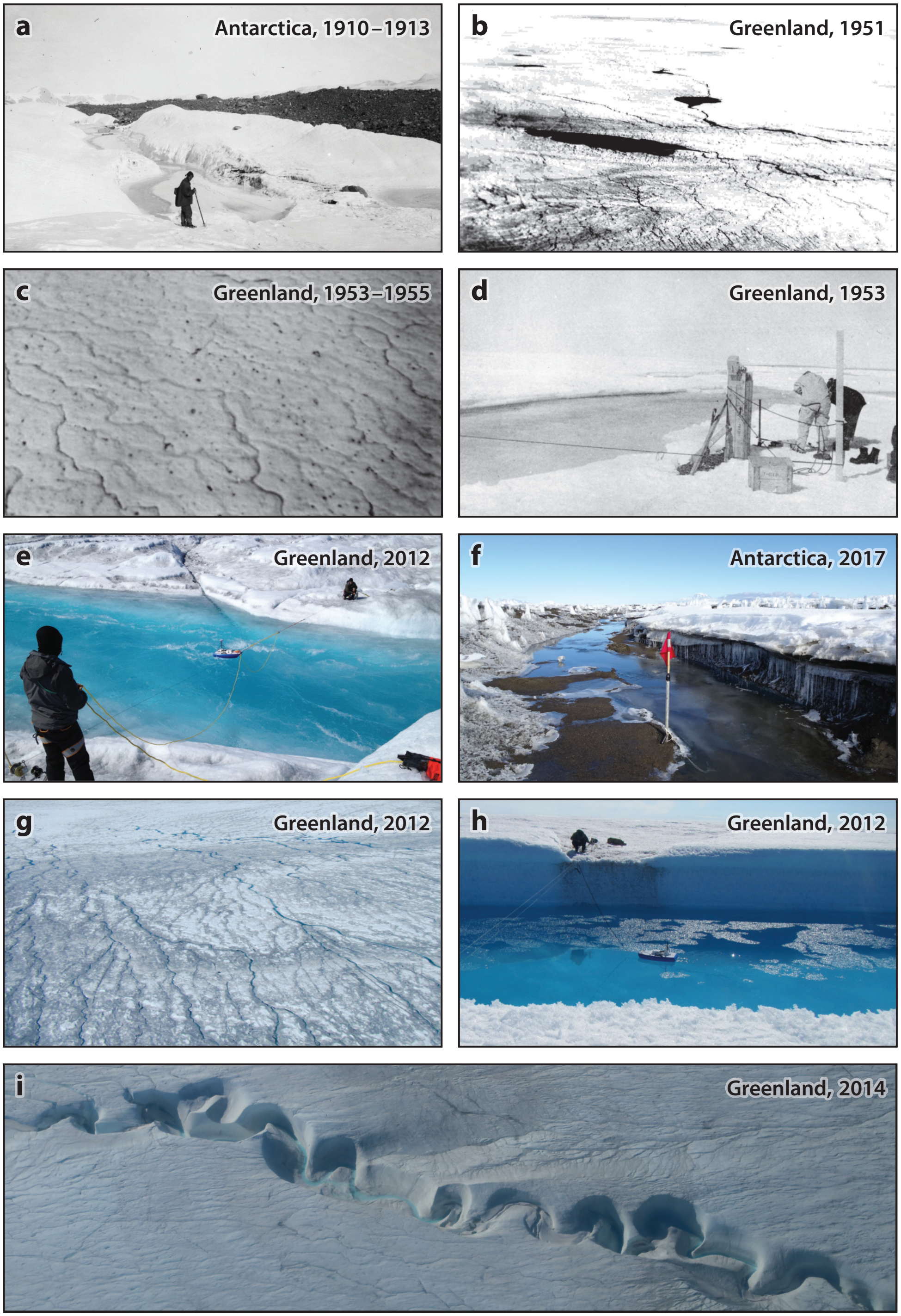
1. INTRODUCTION

Glaciers and ice sheets lose mass via calving and meltwater runoff, which reflect a larger suite of dynamic and hydrologic processes (Rignot et al. 2011, van den Broeke et al. 2009). Appreciation for the importance of hydrological processes continues to grow, owing to climate warming and associated increases in meltwater production across the cryosphere. Furthermore, because the penetration of meltwater can both warm glaciers and alter their basal properties, there are complex feedbacks between meltwater production and long-term stability (e.g., Banwell et al. 2013a; Bell et al. 2017; Colgan et al. 2011; Kingslake et al. 2017; Palmer et al. 2011; Scambos et al. 2004, 2009). Despite this recognition, supraglacial production, storage, and transport on ice masses "remains one of the least-studied hydrologic processes on Earth" (Smith et al. 2015, p. 1001).

The historical record of meltwater transport in supraglacial channels on ice sheets dates to the 1800s. During an early traverse of Greenland, Fridtjof Nansen (1906) recounted crossing channels with depths submerging explorers' ankles. Reports from the British Antarctic Expedition of 1907–1909 led by Sir E.H. Shackleton similarly noted the presence of channelized meltwater in Antarctica (David & Priestley 1914) (**Figure 1*a***). In 1909 a Swiss team led by Alfred de Quervain investigated outlet glaciers in Greenland and noted that meltwater streams made travel difficult. Three years later, Quervain led an eastward traverse across Greenland and recounted that a team member slipped into a meltwater channel upstream of a terminal moulin, regained his footing using crampons, and safely escaped, losing only an ice axe (Barr 2015). In 1930–1931 the British Arctic Air-Route Expedition noted navigating channels in southwest Greenland (Chapman 1932), while a separate 1934 British-led traverse also encountered a handful of rapidly flowing meltwater channels (Lindsay 1935). There are similar reports of supraglacial meltwater channels in Svalbard (Ahlmann & Rosenbaum 1933, Glen 1941, KSS 1934), Scandinavia (Ahlmann 1922, 1923; Lindskog 1928), and the Yukon, Canada (Sharp 1947). Aerial photography enabled surveying ice masses from above, including photointerpretation of meltwater channels in Antarctica (Roscoe 1952).

In the nineteenth century, the US armed forces grew interested in supraglacial channels in the context of aircraft operations and engineering projects on ice. In 1947 the US Army Corps of Engineers initiated Project Snowman, in which a team was deployed to southwest Greenland to study the feasibility of landing aircraft on ice and subsequently identified a complex network of meltwater lakes and channels as an obstacle (USACE 1947) (**Figure 1*b***). In 1951 the United States launched Operation Skyline to develop helicopter search and rescue operations in Greenland and similarly noted an abundance of meltwater-filled lakes and channels emanating inland from the ice edge (Locker 1951) (**Figure 1*b***). The first known discharge measurements in Greenland meltwater channels were made in 1953 as part of the US Army Corps' Project Mint Julep (Holmes 1955, USACE 1953) (**Figure 1*d***). Between 1953 and 1955, the US Army also investigated ice conditions near Thule Air Base in northwest Greenland (**Figure 1*c***). This work further characterized supraglacial meltwater channels, making note of diurnal flow variations, multiyear stability, morphometry, and lateral spacing (Nobles 1960).

Observations by early explorers, coupled with military-led research, set the stage for targeted field investigations, primarily of small channels along the periphery of alpine glaciers. These included examinations of channel morphology, hydraulics, meandering or sinuosity, longitudinal profiles, and analogs with the morphology and processes of terrestrially based bedrock rivers (Dozier 1974, 1976; Ferguson 1973; Knighton 1972, 1981, 1985; Marston 1983; Parker 1975; Sharp 1947; Zeller 1967). Collectively these process-based studies dominated the course of study through the late 1980s (Gleason et al. 2016).



(Caption appears on following page)

Figure 1 (*Figure appears on preceding page*)

Examples of supraglacial stream and river research and process. (*a*) An early polar explorer, a member of the 1910–1913 British Antarctic Expedition, standing next to a supraglacial river in Antarctica. Photo by Frank Debenham. (*b–d*) Aerial photographs from US Armed Forces investigations in the early to mid-1900s of supraglacial channels in the context of aircraft operations and engineering on ice. (*b*) A photograph collected by the Operation Skyline team (scale unknown) in southwest Greenland. Photo taken from Locker (1951). (*c*) A photograph collected as part of the US Army investigation into ice conditions (scale unknown) near Thule Air Base in northwest Greenland. Photo taken from Nobles (1960). (*d*) Members of the 1953 US Army's Project Mint Julep measuring streamflow in southwest Greenland. Photo taken from USACE (1953). (*e*) An example of a streamflow measurement in a supraglacial river in Greenland as part of ongoing research aimed at validating melt and runoff models. Photo provided by Laurence C. Smith. (*f*) An example of a supraglacial river meandering across the McMurdo Ice Shelf. These rivers have been studied in Antarctica for their role in ice shelf stability. Photo provided by Grant Macdonald. (*g*) A dendritic network of supraglacial channels in Greenland (scale unknown). Photo by Lincoln H Pitcher and Laurence C. Smith. (*h*) Vertical channel incision and floating ice crystals or slush (researcher on far bank for scale). Photo by Lincoln H Pitcher. (*i*) A sinuous and incised supraglacial river in Greenland (scale unknown). Photo by Lincoln H Pitcher.

More recently, a heightened awareness of accelerating mass loss trends from glaciers and ice sheets (e.g., Vaughan et al. 2013) has motivated new research on supraglacial meltwater channels, primarily due to their occurrence as visible and integral elements of the hydrological system governing surface mass balance (SMB) (e.g., Smith et al. 2015, 2017), ice dynamics (e.g., Karlstrom et al. 2014), and ice shelf stability (Bell et al. 2017, Kingslake et al. 2017, Macdonald et al. 2018). This has been facilitated by advances in remote sensing that enable manual and automated detection of supraglacial meltwater channels in visible and near-infrared airborne and spaceborne imagery (e.g., Bell et al. 2017; Brykala 1998a; Ewing 1970; Holmes 1955; Kingslake et al. 2017; Lampkin & VanderBerg 2014; Legleiter et al. 2014; Orheim & Lucchitta 1987; Ryan et al. 2016, 2017a,b; Smith et al. 2015; Swithinbank 1988; USACE 1953; Yang & Smith 2013, 2016; Yang et al. 2016a,b) as well as radar data (Munneke et al. 2018, Phillips 1998). Yet the location of meltwater channels in cold, harsh, remote, costly, and logistically difficult-to-study areas has resulted in limited field observations, particularly of large channels on the interiors of ice sheets (Gleason et al. 2016).

There are several reviews of glacier hydrology research that discuss surface meltwater transport. Lawson (1993) focused on meltwater and sediment supply to downstream lake and river systems. Fountain & Walder (1998) proposed a framework for meltwater transport on, in, under, and from glaciers. Irvine-Fynn et al. (2011) detailed hydrologic processes in temperate glaciers with a specific focus on valley glaciers in the Arctic. Greenwood et al. (2016) linked contemporary research with paleo-reconstructions of ice sheet hydrology. Rennermalm et al. (2013a), Chu (2014), Yang & Li (2014), and Flowers (2018) provided overviews of the hydrology of the Greenland Ice Sheet, while Liestøl (1993), Hagen et al. (1993), and Hodgkins (1997) reviewed hydrologic and glaciologic research in Svalbard. There are also reviews that focused on englacial and/or subglacial hydrology (e.g., Hooke 1989, Hubbard & Nienow 1997), while Cuffey & Paterson (2010) covered the role of meltwater in mass balance, ice dynamics, and ice shelf processes more broadly. However, at the time of writing, a review specific to supraglacial meltwater channels is lacking. To that end, this article reviews their (*a*) known spatial distribution; (*b*) formation, evolution, morphology, and sediment transport processes; (*c*) hydrology, hydraulic geometry, hydraulics, and open-channel flow characteristics; and (*d*) impact on surface energy balance, heat transfer, significance for ice sheet basal conditions in Greenland, and ice shelf stability in Antarctica. It concludes with a discussion of knowledge gaps and recommends future research directions.

2. DEFINITIONS AND SPATIAL DISTRIBUTION

2.1. Definitions

A channel transporting meltwater on the surface of an ice mass can be categorized as a supraglacial river or a supraglacial stream (Smith et al. 2015). Channels is also used as a generic term for supraglacial streams and rivers collectively. Supraglacial rivers are primarily main-stem channels with high stream orders that are perennially occupied. They are regularly spaced, form parallel to ice flow directions, have elongated patterns, and often terminate in moulins. In contrast, supraglacial streams are low order with shallow depths, are annual or transient on multiyear timescales, and are often tributary to larger rivers (Ewing 1970, Smith et al. 2015). For example, **Figure 2*e*** shows supraglacial streams emanating from a slush field in Novaya Zemlya, Russia, while **Figure 2*d*** features a small reach of perhaps the longest (Pelto 2018) supraglacial river in Greenland on Nioghalvsfjerdsfjorden Glacier. This review emphasizes larger, perennially active supraglacial rivers because they remain understudied despite their importance for surface energy balance, thermal heat transfer, and meltwater distribution, especially in comparison to supraglacial streams.

Supraglacial channels can terminate in moulins, crevasses, or meltwater lakes/ponds or can drain directly off the ice into terrestrial hydrologic systems or the ocean (**Figure 3**). **Figure 2*a*** shows a network of supraglacial streams that flow tributary to a supraglacial river terminating in a moulin. This network provides an example of an internally drained catchment (IDC), which delimits the ice surface source area for meltwater delivered to a terminal outlet moulin or endorheic lake (Yang & Smith 2016).

2.2. Spatial Distribution

Supraglacial streams and rivers activate during the summer melt season across the cryosphere. A large fraction of supraglacial river research has been conducted on the west coast of Greenland where the ablation zone is dominated by supraglacial channels. In this region, these features have been surveyed in situ (Cowton et al. 2013; Gleason et al. 2016; Holmes 1955; Locker 1951; Smith et al. 2015, 2017; USACE 1947, 1953), observed using remote sensing (Charalampidis et al. 2016; King et al. 2016; Legleiter et al. 2014; Machguth et al. 2016; Ryan et al. 2016, 2017a, 2018; Yang & Smith 2013, 2016; Yang et al. 2015, 2016a,b, 2018), and inferred with models (Clason et al. 2015, de Fleurian et al. 2016, Karlstrom & Yang 2016, Palmer et al. 2011). Further south, near Nuuk, supraglacial rivers have been mapped with remote sensing (Thomsen 1986). Further north, in proximity to Ilulissat, supraglacial rivers have also been studied in situ (Echelmeyer & Harrison 1990, McGrath et al. 2011, Tedesco et al. 2013), with remote sensing (Colgan et al. 2011; Lampkin & VanderBerg 2014; Thomsen 1986; Thomsen et al. 1988, 1989), and using model-based approaches (Arnold et al. 2014; Banwell et al. 2012, 2013b, 2016; Kingslake et al. 2015; Phillips et al. 2011). A handful of studies investigated supraglacial rivers in other regions of Greenland (Bell et al. 2017, Bøggild et al. 2010, Carver et al. 1994, Cathles et al. 2011, Colgan et al. 2015, Macdonald et al. 2018, Nobles 1960, Pelto 2018) and surrounding glaciers and ice caps (Mernild et al. 2006, Sugiyama et al. 2014). Supraglacial streams and rivers have also been studied in Antarctica (Bell et al. 2017; Birnie & Gordon 1980; Fortner et al. 2005; Kingslake et al. 2015, 2017; Orheim & Lucchitta 1987; Phillips 1998; Rack & Rott 2004; Roscoe 1952; SanClements et al. 2017; Swithinbank 1988; Winther et al. 1996), yet in comparison to Greenland, Antarctica has received considerably less attention.

In Europe supraglacial channels have been studied in Austria (Behrens et al. 1971, Burkimsher 1983), Iceland (Dowdeswell 1982, Flett et al. 2017, Jarosch & Gudmundsson 2012, MacDonald

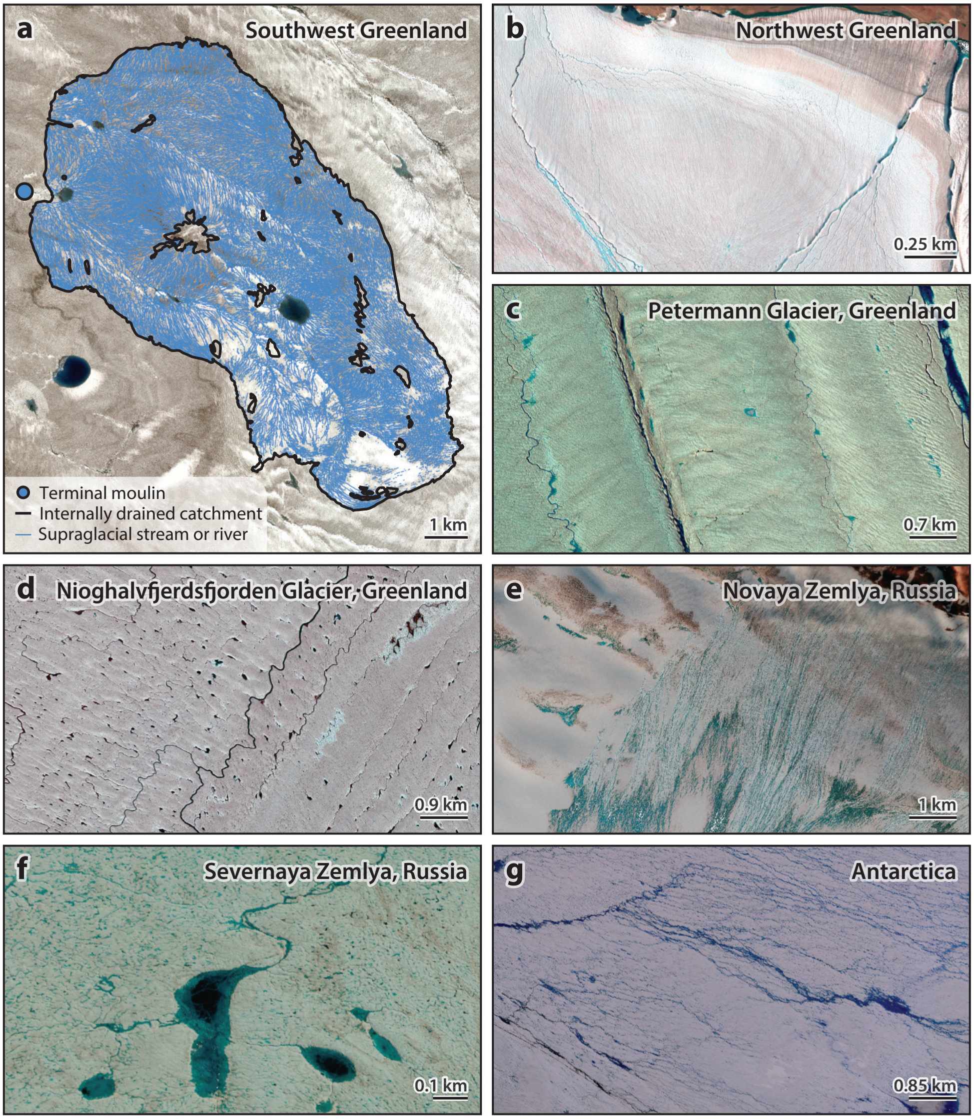


Figure 2

Examples of supraglacial stream and river networks in the Northern and Southern Hemispheres. (*a*) An internally drained catchment (*black*) and supraglacial stream/river network (*blue*). The internally drained catchment boundary and channel networks are from Smith et al. (2017). The image is from WorldView-3, collected July 17, 2015. (*b*) Two incised, land-terminating supraglacial rivers in northwest Greenland. The image is from WorldView-2, collected July 28, 2017. (*c*,*d*) Supraglacial streams and rivers on ice shelves in Greenland. Both images are from WorldView-2, collected July 9, 2015, and July 26, 2013, respectively. (*e*,*f*) Supraglacial streams and rivers on grounded and floating ice in Novaya Zemlya and Severnaya Zemlya, Russia, respectively. Both images are from WorldView-2/-3, collected July 12, 2016, and July 12, 2013, respectively. (*g*) A dendritic supraglacial stream and river network on grounded ice in Antarctica. The image is from QuickBird2, collected January 23, 2011. All images © 2018 DigitalGlobe, Inc.

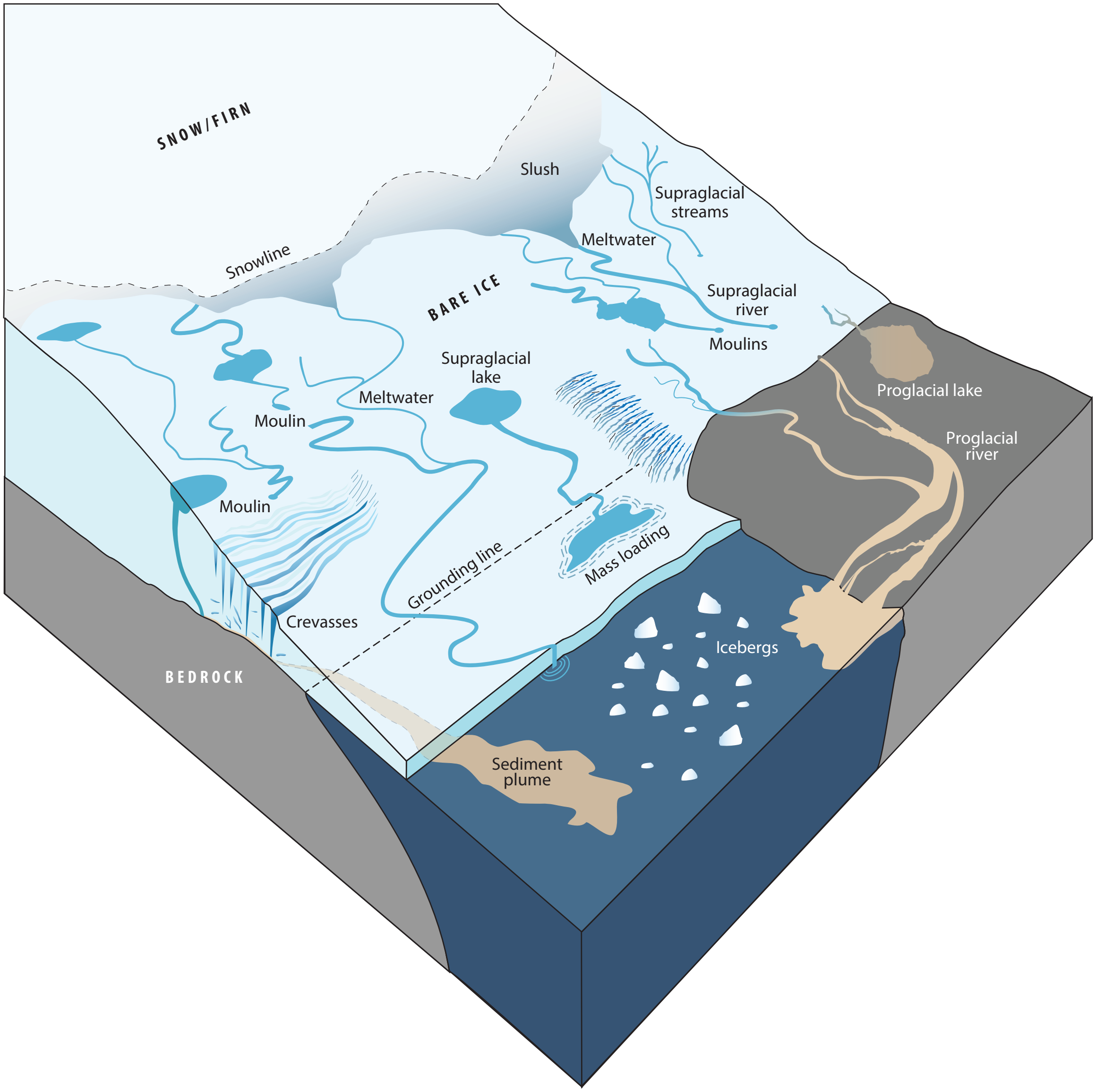


Figure 3

Schematic of supraglacial stream and river processes. Meltwater from ablating snow, firn, and bare ice is transported through supraglacial stream/river channels across bare ice and ice shelves. Their thermally incised channels may terminate directly into moulins, in supraglacial lakes that in turn drain into moulins or outlet streams, or in crevasses. Lake drainages, moulins, and crevasses connect surface climatology and meltwater runoff with en-/subglacial conduits and can modulate a glacier's internal thermal properties, subglacial water pressure, and sometimes sliding velocity. Alternatively, supraglacial streams/rivers may drain directly off the ice into a proglacial lake, river, or fjord with no modification from en-/subglacial processes or remain stored in supraglacial lakes through the winter. In temperate glaciers, moulins commonly supply water to sediment-rich proglacial rivers that emerge where subglacial eskers arrive at the ice edge. In marine environments, esker flows commonly produce turbid sediment plumes in coastal waters near the ice front. Supraglacial streams and rivers can terminate in lakes on ice shelves, which may stress ice shelf integrity via mass loading, or drain directly into the ocean with no impoundment or ice shelf weakening. Illustration by Matt Zebrowski, UCLA Department of Geography.

et al. 2016), Italy (Mantelli et al. 2015), Norway (Hambrey 1977; Knighton 1972, 1981, 1985; Lindskog 1928; Willis et al. 1990), Svalbard (Brykala 1998a,b, 1999; Hagen et al. 1993; Hodgkins 2001; Hodson et al. 2007; Jarosch & Gudmundsson 2012; Kostrzewski & Zwolinski 1995; Rippin et al. 2015), Sweden (Kohler 1995; Seaberg et al. 1988; Stenborg 1968, 1969), and Switzerland (Arnold et al. 1998, Ferguson 1973, Hock et al. 1999, Iken & Bindschadler 1986, Willis et al. 2002, Zeller 1967). There are sparser accounts of supraglacial rivers in Asia (Jarosch & Gudmundsson 2012, Xiao-bo 2018), South America (Isenko et al. 2005), and Russia (Isenko & Mavlyudov 2002, Isenko et al. 2005).

In the United States supraglacial channels have been studied in Alaska (Dozier 1974, 1976; Karlstrom et al. 2013; Marston 1983; Raymond & Nolan 2000; Scott et al. 2010; Sturm & Cosgrove 1990), Washington (Krimmel et al. 1972), and Wyoming (Leopold & Wolman 1960). In Canada supraglacial channels have been studied in Alberta (Hammer & Smith 1983; Mantelli et al. 2015; Munro 2010, 2011), British Columbia (Karlstrom et al. 2014), Nunavut (Bingham et al. 2005, Germain & Moorman 2016, Müller & Iken 1973, Parker 1975, Whitehead et al. 2013), the Northwest Territories (Iken 1972), and the Yukon (Dewart 1966, Ewing 1970, Jarvis & Clarke 1974, Sharp 1947, Stanley 1972).

There is also a lengthy history of injecting moulin-terminating supraglacial channels with tracers to monitor transport times and chemical concentrations exported at a glacier portal. This helps infer the configuration, geometry, and efficiency of en-/subglacial hydrologic networks. Examples of tracer studies that note the presence of supraglacial channels include those conducted in Austria (Behrens et al. 1971, Burkimsher 1983), Canada (Bingham et al. 2005), Greenland (Cowton et al. 2013), Norway (Willis et al. 1990), Sweden (Kohler 1995, Seaberg et al. 1988, Stenborg 1969), Switzerland (Hock et al. 1999), and the United States (Krimmel et al. 1972).

This overview is conservative because it does not review investigations into glacier hydrology written in languages other than English. Similarly, many conference proceedings and master and doctoral theses are not included.

3. FORMATION, EVOLUTION, MORPHOLOGY, AND SEDIMENT TRANSPORT

3.1. Formation and Evolution

During the transition from winter accumulation to the early melt season, ablation zones remain snow covered. As days grow longer, solar heating warms the ice/snow surface and produces meltwater (Kingslake et al. 2015, 2017; Phillips 1998). Meltwater then percolates through underlying snow and ice, refreezes, and establishes impermeable superimposed ice lenses (Benson 1960, Hambrey 1977, Irvine-Fynn et al. 2011). As production increases, inefficient surface drainage with transport rates of <3 mm h^{-1} ensue (Cuffey & Paterson 2010, Fountain & Walder 1998, Irvine-Fynn et al. 2011). When surface snow and ice reach saturation, drainage begins to mobilize the transport of snow and firn, creating slush flow through topographic lows (e.g., Irvine-Fynn et al. 2011, Onesti 1985, Onesti & Hestnes 1989). Fountain & Walder (1998) suggested that supraglacial meltwater channels indicate that near-surface glacier ice is impermeable. Therefore, runoff begins when the near-surface ice becomes saturated, which is similar to hillslope processes for terrestrial systems (Cuffey & Paterson 2010).

The genesis of supraglacial channels remains poorly understood (Irvine-Fynn et al. 2011, Mantelli et al. 2015), but it is thought to be influenced by the rate of channel incision relative to surface ablation (Marston 1983), weathering crust hydrology, meltwater production, and surface topography (Irvine-Fynn et al. 2011). At the beginning of the melt season, rills develop tributary to larger channels, which often form parallel to ice flow directions (Hambrey 1977, Knighton 1972).

Supraglacial channels form along the path of the steepest flow direction (Mantelli et al. 2015), and streamflow is maintained by both runoff from contributing icescapes (e.g., Munro 2010, 2011) and melting along channel walls (e.g., Knighton 1972).

Supraglacial rivers are an important erosional agent for the ice surface (Birnie & Gordon 1980). Many channels are occupied perennially (Ferguson 1973, Glen 1941, Hambrey 1977) and can adjust rapidly (Dozier 1974, 1976; Ferguson 1973; Karlstrom et al. 2014) in response to internal and external forcing. Internal adjustments are caused by thermal erosion and result in geometric modifications such as changes in channel width or depth. External adjustments are forced by glacier flow (Dozier 1974) or by interactions of glacier flow with bedrock topography (Karlstrom & Yang 2016). In general, as the melt season progresses, supraglacial rivers evolve into dynamic, complex, sinuous, dendritic, incised systems (Irvine-Fynn et al. 2011). Kostrzewski & Zwolinski (1995) proposed that channel evolution follows three stages. First, channels incise, resulting in large changes in depth relative to width. Second, ablation along channel walls results in lateral expansion, increasing width relative to depth. Third, meandering (Section 3.2.1) initiates as flow velocity responds to modifications in channel roughness.

This conceptual model is based on internal forcing, but supraglacial rivers are also affected by external processes. Transport of preexisting channels by ice flow and modification of ice topography as it reacts to bedrock topography can alter gradients, realign drainage patterns, elongate channel networks, and set the location of supraglacial rivers at length scales approximately equal to ice thickness (Ewing 1970, Karlstrom & Yang 2016). Dozier (1974) suggested that, like their bedrock counterparts, supraglacial channels evolve toward an equilibrium state and that sinuous systems (Section 3.2.1) are closest to achieving it. Fountain & Walder (1998) disputed this, arguing that an equilibrium state is unattainable due to variations in meltwater supply. Karlstrom & Yang (2016) hypothesized that an equilibrium state is possible if channel incision rates equal surface ablation rates but also hypothesized that such conditions are likely rare. Prevailing theory concludes that at length scales much smaller than one ice thickness, internal thermal adjustments control fluvial channel topography. But at scales approximately equal to ice thickness or more, ablation zone fluvial landscapes are controlled by subglacial topography (Karlstrom & Yang 2016).

3.2. Morphology

The primary erosional process influencing supraglacial channel morphology is the melting of channels as driven by two energetic sources: frictional dissipation of heat as flowing water loses potential energy and energy input to the surface of the channel boundary. The latter is enhanced by lower albedos relative to glacier ice (Section 5.1), topographic shading resulting in variable melt rates, and tributary in- and outflows that modify flow conditions (Karlstrom et al. 2013). Additionally, morphology is influenced by extensional and compressional ice flow (e.g., Karlstrom et al. 2013, Marston 1983). Collectively, these processes govern the morphology of supraglacial channels.

3.2.1. Meandering. It is widely observed that supraglacial channels can meander (Dozier 1974, 1976; Ferguson 1973; Hambrey 1977; Karlstrom et al. 2013; Knighton 1972; Leopold & Wolman 1960; Marston 1983; Parker 1975; Rippin et al. 2015; Zeller 1967) (e.g., **Figures 1*a*,*i*** and **2*c*,*d*,*f***). It is also understood that curvature in terrestrial and supraglacial systems is amplified by both flow against channel boundaries and channel curvature itself, wherein flow either mechanically or thermally erodes the bank. Prevailing theories attribute meandering in alluvial rivers to sediment erosion and deposition (e.g., Braudrick et al. 2009, Church 2006). But meandering in supraglacial channels that lack similar sediment processes yet exhibit similar meander geometries (Leopold & Wolman 1960) suggests that sediment transport is not a requirement for meandering.

There are numerical models that simulate supraglacial channel meandering based on thermal forcing. For example, Parker (1975) proposed that meandering can be explained by thermal erosion due to differential frictional heating around bends. He found that to initiate meandering, inertial forces must be sufficiently large relative to resisting forces (specifically flow must be supercritical; Section 4.3), and that meanders do not migrate downstream. Marston (1983) agreed that differential frictional heating promotes meandering yet disagreed that meander formation requires supercritical flow and that meanders do not migrate downstream. Karlstrom et al. (2013) built upon the Parker (1975) framework by coupling a streamflow model with an ice melt model that accounted for spatially varying frictional heating. Their simulations revealed that meanders can form in subcritical flows when Froude numbers (Section 4.3) exceed ~0.4 and when channel width-to-depth ratios are 2.5 to 5. They concluded that meandering is initiated by channel curvature, which establishes a flow instability triggering differential heat transfer and therefore uneven melt rates along channel walls.

Knowledge of meandering processes in supraglacial channels has progressed since the behavior was first observed. While Karlstrom et al.'s (2013) coupling of a flow model with an ice melt model was a significant advance, their framework assumed a constant channel geometry and did not account for surface melt or non-channelized flow that drives meltwater supply. Model parameters (e.g., ice/water temperatures) were tuned to reproduce field-observed meander geometries and thereby made inferences about natural form and process. To further advance understanding of supraglacial channel meandering, the coupled hydrologic flow and ice melt modeling framework of Karlstrom et al. (2013) should be integrated with temporally evolving channel geometry and realistic discharge and temperature fluctuations constrained by field measurements.

3.2.2. Incision. Supraglacial channels erode laterally and incise vertically (**Figure 1*h*,*i***). Vertical channel incision rates of ~2–4 cm per day and ~4–6 cm per day have been measured in Greenland (McGrath et al. 2011) and Alaska (Marston 1983), respectively, while rates of >10 cm per day have been measured on mountain glaciers in Switzerland, Patagonia, and Russia (Ferguson 1973, Isenko et al. 2005). Examples of modeled and field-measured incision rates are given in **Supplemental Table 1**. Supraglacial channel incision is thermally driven (Ferguson 1973, Fountain & Walder 1998, Isenko & Mavlyudov 2002, Isenko et al. 2005, Kingslake et al. 2015, Marston 1983). Potentially, suspended ice particles (**Figure 1*b***) cause mechanical abrasion (Knighton 1981, 1985), but this remains untested.

Supplemental Material >

Researchers have measured in situ and numerically simulated incision (e.g., Ferguson 1973, Fountain & Walder 1998, Holmes 1955, Isenko & Mavlyudov 2002, Isenko et al. 2005, Jarosch & Gudmundsson 2012, Karlstrom & Yang 2016, Karlstrom et al. 2013, Kingslake et al. 2015, Marston 1983, McGrath et al. 2011, Willis et al. 2002). Models assume that, as in the case of meander bend growth, the primary mechanism of vertical incision is the melting of channel boundaries. Again, melting is driven by the sum of frictional heat dissipation and positive net energy flux. Some models include both energy sources as drivers of incision while others include only one. Some models quantify lowering for a channel cross section, and others do so for a point.

Fountain & Walder (1998) proposed a model that assumes all melt is driven by frictional heat dissipation and did not consider energy fluxes at the water surface. They quantified channel incision ($\dot{d}$) for a theoretical semicircle-shaped open conduit that incises without widening as

$$\dot{d} = \frac{1}{2}\left(\frac{\pi}{2n}\right)^{\frac{3}{8}}\left(\frac{\rho_{\mathrm{w}}}{\rho_{\mathrm{i}}}\right)\left(\frac{g}{L}\right)S^{\frac{19}{16}}Q^{\frac{5}{8}}, \quad 1.$$

where n is Manning's roughness (Section 4.3), approximated as 0.1 s $m^{-1/3}$ for a smooth channel, ρ_w is water density (1,000 kg m^{-3}), ρ_i is ice density (~900 kg m^{-3}), g is acceleration due to gravity, L is the latent heat of melting (335 kJ kg^{-1}), S is the water surface slope, and Q is discharge (m^3 s^{-1}). This model is presented in the context of stream capture in which supraglacial channels become en-/subglacial conduits via thermal and dynamical processes (e.g., Fountain & Walder 1998, Irvine-Fynn et al. 2011) (en-/subglacial drainage constitutes its own subfield and is not covered here). In this model, $\dot{d}$ is the same for open and closed channels given that the conduit maintains a free surface. Note that $\dot{d}$ calculates the channel lowering for the bottom of a semicircle conduit, not the conduit sides.

Jarosch & Gudmundsson (2012) proposed a more complicated framework for simulating incision in which they coupled a nonlinear viscous ice dynamics model with an open-channel flow model and a thermal transfer model. They found that incision is most sensitive to heat transfer from turbulent mixing, which varies with discharge. The primary purpose of this framework is similar to Fountain & Walder's (1998) and is to explain englacial conduit formation; therefore, Jarosch & Gudmundsson (2012) considered only the contribution of heat dissipation to channel melting, not surface energy fluxes.

To enhance understanding of lateral supraglacial lake drainages through meltwater channels (e.g., Raymond & Nolan 2000, Sturm & Cosgrove 1990, Winther et al. 1996), Kingslake et al. (2015) modeled incision for supraglacial channels that drain lakes. Change (Δ) in the outlet channel bottom elevation (d_c) over time (t) is calculated as

$$\frac{\Delta d_c}{\Delta t} = -\frac{f_f \rho_w}{8L\rho_i} {v_z}^3, \qquad 2.$$

where f_f is the channel hydraulic roughness [Darcy-Weisbach roughness from Clarke (2003)] and v_z is the surface velocity. For simulations, Kingslake et al. (2015) assumed that $f_f = 0.25$ based on calculations of Manning's roughness coefficient (n) in supraglacial streams as measured by Mernild et al. (2006) and that

$$f_f = \frac{8gn^2}{R_h}, \qquad 3.$$

where R_h is the hydraulic radius or the ratio of a cross-sectional area and wetted perimeter (Clarke 2003). This model assumes that incision occurs only vertically and that melting along channel sides is negligible. Similar to Fountain & Walder (1998) and Jarosch & Gudmundsson (2012), Kingslake et al. (2015) assumed that all incision is due to frictional heat dissipation without consideration of surface energy fluxes.

Karlstrom & Yang (2016) built upon previous frameworks by considering the sum of surface energy fluxes and frictional heat dissipation. They modeled the average rate of incision ($\dot{d}$) along a channel boundary as

$$\dot{d} = -\frac{\theta + \tau_d \bar{v}}{\rho_i L}, \qquad 4.$$

where θ is the energy balance at the free water surface (W m^{-2}), $\bar{v}$ is average velocity (m s^{-1}), and τ_d is the shear stress on the channel bottom calculated as $\tau_d = \rho_i g d S$. This framework assumes uniform flow and that any additional thermal heating due to channel meandering is negligible. The addition of the surface energy flux term with heat dissipation represents a significant departure from previous models and should similarly be considered in future investigations into incision. There are limited field observations for validation of these models, but a comparison of simulations forced with realistic field observations would help determine optimal applications for each of the proposed frameworks.

3.3. Sediment Transport

Sediment in supraglacial rivers can be organic [e.g., mobilized or deposited cryoconite that is grain-like in form and composed of biologically active algae, bacteria, and other particulates (Irvine-Fynn et al. 2011, Takeuchi 2002)] or inorganic (e.g., minerals, aeolian deposits, rockfall debris, volcanic ash, suspended ice crystals). Sediment concentrations are typically low, and direct measurements are limited. That said, sediment accumulation on channel beds can reduce ice albedo, thereby increasing the proportion of absorbed to reflected shortwave radiation and promoting channel melt. This positive feedback remains poorly quantified (Mantelli et al. 2015, Stibal et al. 2012), and further research is needed. Gleason et al. (2016) hypothesized that cryoconite pitting, which is widely observed in supraglacial channels (e.g., Hodson et al. 2007), can increase flow roughness, particularly in a cross-stream direction, consequently impacting stream velocity and hydraulics (Section 4.2). There is also tangential literature about geochemical and isotopic flux through supraglacial channels (Fortner et al. 2005, MacDonald et al. 2016, SanClements et al. 2017, Scott et al. 2010, Tranter et al. 1993, Xiao-bo 2018) and connected downstream watersheds that is not reviewed here.

Lawson (1993) noted that while suspended loads in supraglacial channels can be 0 g L^{-1}, typical concentrations range from 0.05 to 0.4 g L^{-1} and can exceed 60 g L^{-1}, with the highest concentrations found in channels that drain moraines. In >40 samples from two supraglacial streams on Hilda Glacier in Canada, Hammer & Smith (1983) found suspended loads of <0.5 g L^{-1}, with most samples having loads of <0.25 g L^{-1}. Supraglacial streams near volcanoes are also observed to have high sediment loads. During volcanic activity, ash can be deposited on a glacier surface in the accumulation zone, buried by snowfall over subsequent winters, and transferred to the ablation zone, where it is reexposed as sediment veins that can be mobilized and transported (Dowdeswell 1982). Over a 5-day study of nine supraglacial rivers on Sylgjujökull, west Vatnajökull, Iceland, Dowdeswell (1982) observed sediment concentrations ranging from 0.06 to 0.43 g L^{-1} and loads ranging from 0.2 to 32.7 g s^{-1} that were ultimately deposited at lower elevations.

Suspended ice crystals (also referred to as slush) (see **Figure 1*b***) in supraglacial channels are widely observed (e.g., Chu 2014; Gleason et al. 2016; Holmes 1955; Knighton 1981, 1985; Marston 1983) but have received little specific study. Marston (1983) found that slush load is mostly sourced from the non-channelized surface compared to channel boundaries and that loads vary from ~2.5 to ~25 g s^{-1}, which is consistent with the inorganic loads observed by Dowdeswell (1982). Furthermore, for his field site(s) in Alaska, Marston (1983) observed that slush roughly varies with discharge, that loads pulsate, and that channels are supply rather than transport limited. In general, understanding of the effect of slush transport on channel morphology and hydraulics remains preliminary.

4. HYDROGRAPHS, HYDRAULIC GEOMETRY, FLOW RESISTANCE, AND OPEN-CHANNEL FLOW

4.1. Hydrographs

Given the challenges in glacier field research, supraglacial hydrographs are rare (e.g., Marston 1983, Smith et al. 2017) (**Table 1**). Reconstructing hydrographs from a stage-discharge rating curve should be approached carefully (McGrath et al. 2011) due to rapid changes in channel geometry (Holmes 1955, USACE 1953) and may be unsuitable (Smith et al. 2017). Supraglacial

Table 1 Summary of supraglacial stream and river discharge measurements

Source	Location	Time	Minimum discharge (m^3 s^{-1})	Maximum discharge (m^3 s^{-1})	Number of streams	Method and/or notes
Holmes (1955)	Alpha River, Project Mint Julep, southwest Greenland	July 21 – August 15, 1953	0.14	5.11	1	Evidence of wading measurements with mechanical flow meter (inferred from Holmes 1955, photo 5, p. 28).
Leopold & Wolman (1960)	Dinwoody Glacier, Wind River Range, Wyoming, USA	NA	0.07	0.07	>1	NA
Knighton (1972)	Østerdalsisen Glacier, Svartisen ice cap, Norway	July–August, year NA	0.005	0.02	1	NA
Ferguson (1973)	Lower Arolla Glacier, Switzerland	1967	<0.01	>0.08	3	Surveys of channel geometry at 20 stations on three streams plus salt-dilution discharge measurements Range in discharge not specified Approximated from Ferguson (1973, figure 2)
Müller & Iken (1973)	White Glacier, Axel Heiberg Island, Canada	1959–1969	~0	>0.15	NA	Approximated from Müller & Iken (1973, figure 2 and figure 3)
Dozier (1974)	Capps Glacier, Chitistone Pass, Wrangell Mountains, Alaska, USA	1969	0.0076	0.30	2	Manual measurements of stream width and depth Velocity measured with Price-type current meter
Knighton (1981)	Austre Okstindbreen, Norway	NA	0.002	0.052	3	NA
Dowdeswell (1982)	Sylgjujökull, west Vatnajökull, Iceland	July 1979	0.003–0.107 (mean peak discharge, 3-h records for 5 days)	0.003–0.107 (mean peak discharge, 3-h records for 5 days)	9	Product of cross-sectional area and velocity
Echelmeyer & Harrison (1990)	Jakobshavn Isbræ, west Greenland	1985–1986	50–80	50–80	1	Discharge range given for large river flowing along center of glacier that terminates above ~900-m elevation

(*Continued*)

Table 1 (*Continued*)

Source	Location	Time	Minimum discharge (m^3 s^{-1})	Maximum discharge (m^3 s^{-1})	Number of streams	Method and/or notes
Carver et al. (1994)	Harlech Gletscher, east Greenland	August 1989	0	1.0	1	Observations of pulsating discharge with 6- to 7-s lags between pulses
Kostrzewski & Zwolinski (1995)	Ragnarbreen, west Spitsbergen, Norway	July 1985	0.003	0.016	1	Width and depth manually measured Surface velocity measured with float method and correction applied for depth integrated velocity
Kohler (1995)	Storglaciaren, Sweden	1988	~0.25	~1.5	5	Discharge inferred from Kohler (1995, figure 3)
Brykala (1999)	Waldemar Glacier, northwest Spitsbergen, Norway	July–August 1997	0.00005	0.0947	1	NA
Hock et al. (1999)	Aletschgletscher, Switzerland	1990, 1991	0.0005	0.002	Several	NA
Willis et al. (2002)	Haut Glacier d'Arolla, Switzerland	August 1993	~0 ± 0.014	>0.25 ± 0.014	1	Stage-discharge rating curve established using detrended pressure transducer data and six discharge measurements Error given as root mean square error
Mernild et al. (2006)	Mittivakkat Glacier, Ammassalik Island, southeast Greenland	August 2004, May 2005	0.012	0.034	5	Hydraulic radius (product of width and depth) manually measured at cross sections Velocity calculated using Manning's formula
Bingham et al. (2005)	John Evans Glacier, Ellesmere Island, Canada	2000–2001	>0	<2	1	Stage-discharge rating curve
Scott et al. (2010)	Mendenhall Glacier, Alaska, USA	NA	0.01	0.02	1	NA
McGrath et al. (2011)	69.554°N, 49.899°W, west Greenland	August 3–17, 2009	0.18 ± 0.05 (daily average)	0.18 ± 0.05 (daily average)	1	Manual survey of cross-sectional area Stage recordings with noncontact sonic level sensor Velocity measurements of in-channel propeller
Karlstrom et al. (2014)	Llewellyn Glacier, Juneau Ice Field, British Columbia, Canada	August 2010	0.01 (daytime average)	0.01 (daytime average)	>1	Manual measurements of width and depth Velocity measurements with acoustic Doppler velocimeter

(*Continued*)

Table 1 (*Continued*)

Source	Location	Time	Minimum discharge ($m^3\ s^{-1}$)	Maximum discharge ($m^3\ s^{-1}$)	Number of streams	Method and/or notes
Smith et al. (2015)	Southwest Greenland Ice Sheet	July 2012	0.36 ± 3.6	17.72 ± 3.6	523	Width retrieval from high-resolution WorldView-1/-2 satellite imagery Depth retrieval from WorldView-1/-2 using field-calibrated spectral relationship between depth and reflectance (Legleiter et al. 2014) Velocity determined from field-calibrated hydraulic-geometry relationships
Germain & Moorman (2016)	Fountain Glacier, Bylot Island, Nunavut, Canada	2014	0.01	1	1	Stage-discharge rating curve established with 13 manual measurements
Gleason et al. (2016)	Southwest Greenland Ice Sheet	2012	Small: 0.006 Large: 4.58	Small: 0.402 Large: 23.12	9	Small streams surveyed with manual mechanical instruments Large streams surveyed with acoustic Doppler current profiler
Smith et al. (2017)	Southwest Greenland Ice Sheet	July 2015	4.61	26.73	1	Acoustic Doppler current profiler
Bell et al. (2017)	Nansen Ice Shelf, Antarctica	2006–2015	259	806	1	Width calculated from Landsat-8 Depth retrieval using reflectance-depth optical band ratios Velocity estimates using Manning's equation
SanClements et al. (2017)	Cotton Glacier, McMurdo Dry Valleys, Antarctica	2011	0.11 (one measurement)	0.11 (one measurement)	1	Width calculated manually Depth from pressure transducer Hydroacoustic velocity

hydrographs have been inferred from climatology forced-runoff models (e.g., Banwell et al. 2013b, de Fleurian et al. 2016, Willis et al. 2002), yet one study found that models overpredict discharge by 21–58% and do not reproduce the timing of peak discharge (Smith et al. 2017).

The supraglacial river hydrograph is characterized by large diurnal and seasonal variability. Depending on the size and shape of the catchment, peak discharge lags behind peak melt, which tracks peak solar radiation (Dozier 1974; McGrath et al. 2011; Munro 2010, 2011; Smith et al. 2017; Willis et al. 2002). This lag is due to catchment shape and area, with large, elongated catchments having longer delays than short, compact ones (Munro 2010, Smith et al. 2017); residual drainage from weathering crust, a low-density surface that modulates the timing of hillslope processes and meltwater delivery to channels (Cooper et al. 2018, Irvine-Fynn et al. 2011, Karlstrom et al. 2014, Munro 2011, Smith et al. 2017, Willis et al. 2002); and drainage density of the channel network, with high density associated with faster routing (Yang et al. 2018).

A motivation for studying the supraglacial river hydrograph is to constrain the timing and volume of meltwater delivery to moulins, which modulate subglacial water pressure and sometimes basal sliding (e.g., Andrews et al. 2014, Bartholomew et al. 2011a, Chandler et al. 2013, Chu et al. 2016a, Cowton et al. 2013, de Fleurian et al. 2016, McGrath et al. 2011). However, it is unrealistic to monitor discharge in situ for numerous or even one channel for more than a few days. Therefore, routing the output of climatology-based melt/runoff models through a watershed routing model (e.g., Arnold et al. 1998, 2014; Banwell et al. 2012, 2013b, 2016; Clason et al. 2015; de Fleurian et al. 2016; Leeson et al. 2015; Willis et al. 2002) remains the most realistic mechanism for estimating discharge delivered to moulins. To that end, Smith et al. (2017) applied synthetic unit hydrograph (SUH) theory to convert modeled melt into hydrographs at a terminal outlet moulin. The empirical SUH coefficients were derived using 72 continuous hours of in situ discharge measurements collected in Greenland in July 2015. They demonstrated that IDC area, shape, and river length dictate the timing and magnitude of peak meltwater delivery to moulins. Building on this and using the same hydrograph, Yang et al. (2018) partitioned meltwater routing into non-channelized surface (or interfluve) flow and channelized flow. They found that representative interfluve flow distances are 0–100 m compared to 10 km for channelized flow. Smith et al.'s (2017) and Yang et al.'s (2018) findings were calibrated for one IDC and one 3-day snapshot, yet the studies provided a first set of empirical measurements that enabled modeling of moulin hydrographs using classical unit hydrograph and hillslope transport theory.

4.2. Hydraulic Geometry and Flow Resistance

Hydraulic geometry (HG) is an empirically derived set of equations that relates changes in channel width (w), depth (d), and velocity (v) to changing discharge (Q), both at a given cross section [at-a-station hydraulic geometry (AHG)] and in a downstream flow direction [downstream hydraulic geometry (DHG)]. HG was introduced by Leopold & Maddock (1953) and has since been widely applied to supraglacial systems (e.g., Brykala 1999, Gleason et al. 2016, Kostrzewski & Zwolinski 1995, Marston 1983). The HG theorem states that for a given cross section, w, d, and v vary as a power function of Q (Leopold & Maddock 1953). The equations governing HG are

$$w = aQ^b,\ d = cQ^f, \text{and}\ v = kQ^m, \tag{5.}$$

where a, b, c, f, k, and m are empirically derived constants. It follows that

$$b + f + m = a \times c \times k = 1, \tag{6.}$$

where b, f, and m are calculated as the slope of the linear regression when w, d, or v is plotted against Q in logarithmic form. While a, c, and k are the intercepts of the same logarithmic regressions and are therefore equivalent to respective values of w, d, and v when $Q = 1$ (Leopold & Maddock 1953), Gleason & Smith (2014) and Gleason & Wang (2015) revealed empirical correlations between AHG intercepts and coefficients for w, d, and v when averaged over long reaches of terrestrial rivers, a phenomenon they term at-many-stations hydraulic geometry (AMHG). While AMHG has been successfully inverted to estimate terrestrial river discharge from remote sensing (Bonnema et al. 2016, Durand et al. 2016, Gleason & Smith 2014, Gleason et al. 2014, Hagemann et al. 2017), this approach has yet to be applied to supraglacial channels.

Gleason et al. (2016) summarized AHG and DHG coefficients for supraglacial research. A primary takeaway of this summary is that at a given cross section, higher Q is accommodated by larger increases in v relative to w and d (Gleason et al. 2016, Knighton 1981, Marston 1983). Similarly, with higher Q, d increases faster than w (Brykala 1998a, Gleason et al. 2016, Kostrzewski & Zwolinski 1995, Marston 1983), which is consistent with observations of vertical channel incision rates exceeding those of lateral channel melting (Marston 1983). In Greenland, Smith et al. (2015) aggregated in situ measurements of w, d, and v from 54 small supraglacial streams and 24 cross sections across three large supraglacial rivers to calibrate an interchannel AHG power-law function relating Q to w ($w = 3.84Q^{0.54}$, $r^2 = 0.89$, with a root mean square error of 3.11 $m^3\ s^{-1}$). The stability of this empirical relationship among the 78 measured cross sections was attributed to incision of similar channel geometries. DHG investigations suggest that supraglacial Q increases downstream, resulting in subsequent increases in w, d, and v (Gleason et al. 2016, Knighton 1981, Marston 1983), which is consistent with DHG in terrestrial rivers (Leopold & Maddock 1953).

4.3. Open-Channel Flow

Open-channel or free-surface flow conditions are dictated by gravity, channel slope, and friction or channel roughness (Knighton 1998). Flow conditions can be categorized as subcritical, critical, or supercritical. With subcritical flow, gravity and friction are in balance and flow depth/velocity remains consistent over short time intervals. In contrast, critical and supercritical flows are characterized by abrupt changes in flow depth/velocity over short time intervals, akin to downstream flood wave propagation (Chow 1959, Knighton 1972). Flow conditions can be numerically defined by thresholding the instantaneously derived Froude number (F_r), a dimensionless quantity calculated as the ratio between channel velocity, gravity, and depth:

$$F_r = \frac{\bar{v}}{\sqrt{gd}}. \qquad 7.$$

Flows with $F_r < 1$ are subcritical, $F_r = 1$ are critical, and $F_r > 1$ are supercritical (e.g., Hugget 2007).

Flows in supraglacial channels are often subcritical but can be supercritical (e.g., Ferguson 1973; Gleason et al. 2016; Knighton 1981, 1985; Marston 1983; Parker 1975). For example, Leopold & Wolman (1960) calculated $F_r = 1.9$ on Dinwoody Glacier, Wyoming. Carver et al. (1994) found evidence of supercritical flow with $F_r > 2$ on Harlech Gletscher, Greenland, when they observed roll waves with 6- to 7-second lags between pulses propagating downstream. This is consistent with Knighton's (1981) observations on Austre Okstinbreen Glacier, Norway, and Germain & Moorman's (2016) observations on Bylot Island, Nunavut. Gleason et al. (2016) calculated supercritical flows in small ($Q < 0.5$ $m^3\ s^{-1}$), steep (slopes > 0.7%) streams with maximum $F_r = 3.11$ in Greenland, while the rivers ($Q = 4.58$ to 23.12 $m^3\ s^{-1}$) were subcritical (minimum $F_r = 0.45$). In general, supraglacial channels display both subcritical and supercritical flow.

An important hydraulic parameter for open-channel flow is resistance, which modulates and inversely correlates with velocity. Flow resistance is approximated as channel roughness using Chezy, Manning's, or Darcy-Weisbach equations [refer to chapter 4 of Knighton (1998) for a review of resistance and roughness calculations]. The most commonly used resistance metric in supraglacial hydrology is Manning's n, which is calculated as

$$n = \frac{1}{\bar{v}} R_{\mathrm{h}}^{2/3} S^{1/2}, \qquad 8.$$

where $\bar{v}$ is the average velocity, R_{h} is the hydraulic radius, and S is the channel slope. Flow resistance is composed of boundary resistance including substrate and form friction, channel resistance including irregularities that disturb flow, and free surface resistance caused by unsteady flow and potentially slush at the water surface (Knighton 1998). Morphologically, meander bends disturb flow and increase roughness, which suggests that variations in roughness might be reflected by variations in sinuosity. From field measurements, supraglacial channels are observed to have a large range in roughness despite having a uniform substrate and limited sediment and being free of alluvial bedforms. Gleason et al. (2016) hypothesized that such variations may be due to extensional and compressional fractures in channels and bank scalloping, cryoconite pitting, slush, and low water temperatures associated with high viscosity. However, Yang et al. (2018) suggested a mean n value of 0.03–0.05 when averaged across the scale of an entire supraglacial stream/river catchment. In situ characterizations of open-channel flow conditions in supraglacial channels of varying sizes, planforms, and discharges are needed for remote estimation (e.g., Kingslake et al. 2017, Smith et al. 2015) of discharge.

5. SURFACE ENERGY BALANCE, HEAT EXCHANGE, SUBGLACIAL CONNECTIVITY IN GREENLAND, AND ICE SHEET/SHELF STABILITY IN ANTARCTICA

Supraglacial channels influence surface melt by modulating surface reflectance (albedo) and topographic roughness. They also impact the thermal regime of ice by descending into a glacier, freezing, and releasing latent heat. Most recently, supraglacial rivers are being investigated for their role in ice shelf stability. These processes of surface energy balance, heat exchange, subglacial connectivity, and ice shelf stability are reviewed in the following subsections.

5.1. Surface Energy Balance

Supraglacial channel initiation and maintenance require surface melt and runoff. Melt is driven by the surface energy balance, particularly net shortwave radiation, which is the difference between downward and reflected shortwave radiation (van den Broeke et al. 2008). Net shortwave radiation is modulated by surface albedo, which is the total reflected shortwave radiation across all wavelengths. Albedos can range from >0.8 for snow to <0.4 for ice (Cuffey & Paterson 2010). Low albedos enhance melt because a higher proportion of downward shortwave radiation is absorbed rather than reflected. Importantly, the albedo of surfaces wetted by meltwater is less than half that of clean ice (Ryan et al. 2018). For example, in Greenland the field-measured albedo of a melt pond at depths of <5 m ranges from ~0.15 to 0.3 (Tedesco & Steiner 2011). Remotely sensed albedo in supraglacial channels ranges from 0.16 to 0.26 (Ryan et al. 2016, 2017a), while average clean ice albedo is ~0.55 (Ryan et al. 2018). Furthermore, low albedos correlate with high melt (e.g., Greuell 2000), while 15% of albedo variability in west Greenland can be explained by the presence of supraglacial meltwater (Ryan et al. 2018). Cryoconite deposition and pitting are also present in many streams and rivers, which will further reduce albedo (Gleason et al. 2016). Collectively, this suggests that supraglacial channels help modulate the spatial distribution of albedo as they transport sediment (Section 3.3) and meltwater from saturated ice to slush and to ponds and eventually excavate it from the surface.

Supraglacial channels also alter the topography of the ice surface via vertical and lateral channel incision (Section 3.2). Local surface topography modulates the spatial variability of melt, especially in the cross-glacier direction (Arnold et al. 2006). That is, deviation from a flat surface and the subsequent introduction of variable topography establish a rougher surface that can intersect and absorb incoming and reflected radiation. Cathles et al. (2011) modeled the feedback between surface roughness, albedo, and ablation, finding that topographic features, simulated as circular and V-shaped canyons, enhance melt at local scales and therefore influence surface meltwater transport (Cathles et al. 2011). Rippin et al. (2015) investigated linkages between channel presence and surface reflectance (as a proxy for albedo), finding that high channel densities correlate with low reflectance—confirming an important feedback between channel-induced topography, albedo, and melt.

The presence and evolution of supraglacial channels also affect non-channelized icescapes. Vertical channel incision creates relief that propagates as non-channelized flow and small slush-filled rills, and tributary streams develop on ablating slopes tributary to valley floor channels. This channel propagation alters surface topography and albedo. The importance of fluvially eroded icescapes due to channel evolution for surface energy balance has received little study.

Collectively, this suggests that the future expansion of meltwater under a warming climate (e.g., Howat et al. 2013, Ignéczi et al. 2016, Leeson et al. 2015) will further lower ice surface albedos, reducing the fraction of reflected shortwave radiation and promoting melt. It is likely that such an albedo reduction will be attributable to both the expansion of surface meltwater and related propagation of meltwater channels that dissect the ice surface and produce localized topographic features. This also suggests a self-reinforcing feedback between positive net shortwave radiation, melt production, and supraglacial channel evolution.

5.2. Heat Exchange

Supraglacial channels deliver meltwater to internal and subglacial hydrological networks that convey it through, beneath, and beyond a glacier (Fountain & Walder 1998). A large portion of this meltwater is rapidly exported via subglacial meltwater portals (van As et al. 2017). However, some fraction (Chu et al. 2016b, Hodgkins 2001, Rennermalm et al. 2013b, Smith et al. 2015) is retained within the ice, where, if it refreezes, it releases latent heat, thus warming the ice. Such warming is in addition to heat transfer from through-flowing water with initial temperatures warmer than the internal temperature of the glacier (Lüthi et al. 2015, Phillips et al. 2010). The warming produced by these processes, often focused at stable moulins and crevasses (Catania & Neumann 2010, Colgan et al. 2011, McGrath et al. 2011), is collectively termed cryo-hydrologic warming (Phillips et al. 2010). This process was investigated as a mechanism for the thermal response of ice masses to surface melt and has been documented in both Greenland (Charalampidis et al. 2016; Harrington et al. 2015; Lüthi et al. 2015; Phillips et al. 2010, 2013) and the Steele Glacier, Yukon, Canada (Jarvis & Clarke 1974).

Cryo-hydrologic warming can explain differences between field-measured and modeled ice temperature profiles (Harrington et al. 2015, Lüthi et al. 2015). Phillips et al. (2010) found that cryo-hydrologic warming can occur on timescales of years to decades and is modulated by the horizontal spacing of the hydrologic system. This suggests that supraglacial meltwater volume and channel spacing influence cryo-hydrologic warming. It was also proposed that cryo-hydrologic warming enhances ice velocity by warming basal temperatures and reducing viscosity (Phillips et al. 2013). Furthermore, models simulate that the onset of abundant surface melt reaching the bed near the interior of the Greenland Ice Sheet would cryo-hydrologically warm the ice, reduce its viscosity, and promote a thermal-viscous collapse over several thousand years (Colgan et al. 2015). This stresses the importance of water and heat transfer delivered by supraglacial channels to en-/subglacial networks.

5.3. Subglacial Connectivity in Greenland

In Greenland, particularly along the western land terminating flank of the ice sheet, a complex network of supraglacial rivers transports large volumes of surface meltwater to moulins, lakes, and crevasses (Colgan et al. 2011, Smith et al. 2015). Surface melting, discharge, and the spatial location of supraglacial rivers positively correlate with ice velocity (Bartholomew et al. 2011b, Palmer et al. 2011) as well as broadscale bedrock topography (Yang et al. 2015). The mechanisms by which meltwater transported by supraglacial rivers affect dynamical processes differ if the river is terminated by a moulin, crevasse, or lake.

Moulin-terminating rivers inject meltwater directly into en-/subglacial drainage systems. The timing of meltwater delivery to the moulin is determined by its contributing supraglacial river transport capacity, surface climatology, and IDC properties (Smith et al. 2015, 2017; Yang & Smith 2016; Yang et al. 2018). If moulin discharge exceeds the subglacial drainage system capacity, then subglacial water pressures increase and basal sliding may ensue. It is hypothesized that eventually the subglacial drainage system adapts to moulin discharge and basal sliding ceases, while if the moulin discharge is less than the subglacial drainage system capacity, there is little change in water pressure or basal sliding (e.g., Colgan et al. 2011, Schoof 2010).

Crevasse-terminating channels often deliver meltwater to a distributed englacial system. If moulins form from water-filled crevasses, the same subglacial water pressure and ice velocity feedback is invoked. Alternatively, crevasses can diffuse flow, resulting in low-amplitude hydrograph delivery to the subglacial system, a minimal change in subglacial water pressure, and sliding (Colgan et al. 2011).

Supraglacial river discharge entering lakes can be temporarily impounded before escaping at a lake outlet (e.g., Smith et al. 2015) or draining directly into the ice via a hydrofractured moulin that opens at the lake bottom (e.g., Das et al. 2008, Selmes et al. 2011, Tedesco et al. 2013). Therefore, rapid lake drainage events might overwhelm the subglacial drainage system capacity and enhance sliding, while slowly draining lakes and those having river outlets may result in a lower amplitude injection to the bed having minimal impact on sliding (e.g., Das et al. 2008, Tedesco et al. 2013). Meltwater can also be retained in lakes, with their surfaces freezing over during winter (Koenig et al. 2015). Such lakes should have no significant impact on basal water pressures and sliding.

5.4. Stability of Antarctic Ice Sheet and Shelves

The deformation of ice shelves past embayment walls and over basal islands provides a drag force that restrains the seaward flow of marine-terminating glaciers and ice streams. Thus, ice shelf disintegration can cause large increases in ice velocity, upstream ice thinning, and sea level rise. Meltwater on ice shelves is linked with ice shelf stability (Rack & Rott 2004, Scambos et al. 2000), while supraglacial rivers may strengthen (Bell et al. 2017) and weaken (Kingslake et al. 2017) shelf integrity.

Kingslake et al. (2017) outlined a weakening mechanism in which supraglacial rivers can promote ice shelf collapse by filling lakes, which increases the load on an ice shelf and can result in bulging and fracturing (Banwell et al. 2013a, Scambos et al. 2009) (**Figure 3**). Lakes also ablate faster than surrounding ice due to lower albedos (Section 5.1) and can hydrofracture or drain, resulting in shelf rebound. This cycle of loading, fracture, drainage, and rebound induces structural weakness that can cause widespread lake drainage and "a fracture system capable of driving explosive [ice shelf] breakup" (Banwell et al. 2013a, p. 5872). In contrast, Bell et al. (2017) proposed that supraglacial rivers may buffer ice shelves from collapse by efficiently transporting mass off ice shelves into the ocean without impoundment/loading (**Figure 3**). The differing impacts that supraglacial rivers may have on ice shelf stability underscore the need for further study.

6. RESEARCH FRONTIERS

There remain gaps in understanding of supraglacial processes. For example, supraglacial channels are well studied in Greenland, yet those in Antarctica have received less attention. Morphologically, vertical channel incision is noted in the field and complex modeling frameworks have been developed, yet a comparison of incision models forced with field observations is lacking. Ice crystal transport in channels is well documented, yet understanding of the effect of slush on channel morphology and hydraulics remains preliminary. Additionally, four outstanding research questions with relevance to climate variability and global sea level rise are highlighted in the following sections.

6.1. What Is the Geographical Distribution of Supraglacial Rivers Globally, and How Might It Change in the Future?

The global distribution of supraglacial rivers remains unmapped, which is expected given that they are seasonally ephemeral and generally located in cold, harsh, inaccessible places. Also, most supraglacial rivers are narrow, which makes them challenging to map using historically available satellite imagery such as ASTER, Landsat, and MODIS with 15-m, 30-m, and 250-m pixel sizes, respectively. However, advances in cloud-based computing power coupled with improved satellite resolutions and novel automated image classification software packages now enable repeat mappings of surface water bodies (Pekel et al. 2016) as well as bedrock and alluvial rivers at continental (Allen & Pavelsky 2015) to global scales (Allen & Pavelsky 2018). Furthermore, high-resolution satellite data (e.g., Sentinel-2/-3 and WorldView-1/-2/-3/-4) have been used to map supraglacial streams and rivers in Greenland (e.g., Karlstrom & Yang 2016; King et al. 2016; Poinar et al. 2015; Yang & Smith 2013, 2016; Yang et al. 2015, 2016a,b) and Antarctica (e.g., Bell et al. 2017, Kingslake et al. 2017), while the recent advent of CubeSats, such as the Planet cluster (e.g., Cooley et al. 2017, 2019), which now images the entire Earth each day, offers the exciting possibility of the high spatial and temporal resolutions needed to capture these fine-scale, dynamic, and often short-lived features. Collectively, this suggests that global supraglacial river mapping is now feasible.

Supraglacial rivers create fluvial fingerprints indicating active meltwater production and transport across the ice surface, and their termination points reveal moulin locations (Smith et al. 2015, Yang & Smith 2016) and surface-to-bed connections. Therefore, the widescale mapping of supraglacial rivers informs not only the presence of melting ice but also where surface climatology interacts with subglacial water pressures and therefore ice dynamics (de Fleurian et al. 2016). In contrast, supraglacial rivers that drain directly off the terminus of an ice mass may indicate cold-based glaciers and locations where ice dynamics receive little or no surface meltwater. A global supraglacial river map is needed to discern where ice masses receive inputs of heat and water from surface meltwater and where they do not. Coupled with topographic information and climate model outputs, such a map could be used to assess future distributions of supraglacial rivers using climate model simulations. This could help inform which regions of the Greenland and Antarctic Ice Sheets and associated ice shelves are most vulnerable to positive feedbacks between climate warming, supraglacial meltwater transport, and potential linkages to the bed.

6.2. How Accurate Are Ice Sheet Runoff Models?

The primary method for forecasting sea level rise contributions from glaciers and ice sheets uses regional or global climate models to estimate SMB and runoff (e.g., van den Broeke et al. 2009). SMB runoff is often validated via comparisons with measured subglacial outflow or discharge in proglacial rivers (e.g., Mernild et al. 2008, 2012; Overeem et al. 2015; Rennermalm et al. 2013b; Smith et al. 2015; van As et al. 2014, 2017). However, surface meltwater expelled via subglacial conduits into proglacial rivers has been routed through a host of en-/subglacial processes and is therefore removed from the SMB process of meltwater production and runoff. Such comparisons of simulated runoff and observed proglacial river discharge have identified dissimilarities of at least 38% (Overeem et al. 2015, Smith et al. 2015). There are also discrepancies between satellite-based gravimetric observations of mass change and SMB models in melt-prone areas of Greenland (Sasgen et al. 2012, Xu et al. 2015). At the IDC scale (**Figure 2*a***), SMB models oversimulate runoff by at least 21% (Smith et al. 2017). Discrepancies between satellite/field observations and models emphasize the need for validation, interrogation, and refinement of SMB modeling. A promising strategy for such research is direct monitoring of supraglacial river flows across space and time. This has implicit societal relevance given that SMB models are used to forecast sea level rise.

6.3. Do Supraglacial Rivers Influence Ice Dynamics?

Current ice sheet models do not consider hydrologic flow paths through channelized or non-channelized ice, nor do they account for point source locations of surface-to-bed hydrologic connections (e.g., moulins). This is noteworthy because ice surface hydrology preconditions the magnitude and timing of meltwater delivery to the bed (Smith et al. 2017) and is therefore an important control on subglacial hydrologic flow gradients, water pressures, and ice sliding (Schoof 2010). Consequently, the lack of coupled hydrologic and ice dynamics models impedes accurate coupling of SMB with ice flow, especially at short (i.e., diurnal) timescales. Therefore, coupling a surface hydrology-routing model with an ice dynamics model that accurately simulates the volume, magnitude, and location of meltwater injection to the bed will improve our understanding of the future response of glaciers and ice sheets to increased meltwater production.

6.4. Does Meltwater Transport on Ice Shelves Inhibit or Accelerate Ice Shelf Collapse?

Geological records indicate that during the Last Interglacial and the Pliocene (~130,000 and ~3 million years ago, respectively), global mean sea levels were >6 m higher than today (Dutton et al. 2015). DeConto & Pollard (2016) explained that most of this historical difference is attributed to a smaller Antarctic Ice Sheet, which motivates the use of coupled ice-ocean-atmosphere models to reconstruct past conditions and assess the vulnerability of the Antarctic Ice Sheet to the current climate. DeConto & Pollard (2016) found that simulating ancient Antarctic Ice Sheet retreat on scales commensurate with geological sea level rise records requires the retreat and collapse of major ice shelves, which currently stabilize and buttress grounded ice. To mimic historical ice shelf retreat and collapse, models require ocean warming, resulting in subsurface melt and shelf thinning, and/or surface meltwater production, resulting in thinning, crevassing, calving, and hydrofracturing. This suggests that ice shelves are vulnerable to meltwater production, which is amplified by low hypsometry whereby small temperature increases yield disproportionately large melt area increases (DeConto & Pollard 2016). As reviewed in this article, the impact of meltwater transport on ice shelf stability has been interpreted as both a stabilizing (Bell et al. 2017, Macdonald et al. 2018) and a destabilizing (Kingslake et al. 2017) process. Further research is needed to enhance understanding of pathways and processes of ice shelf meltwater production, transport, and export to determine how supraglacial channels impact ice shelf stability and therefore global sea level rise.

7. CONCLUSIONS

This work reviews and synthesizes published English-language studies of supraglacial streams and rivers on glaciers, ice sheets, and ice shelves. Supraglacial rivers link surface climatology with ice dynamics and are an important component of how glaciers and ice sheets respond to climate variability. Despite this, relatively little is known about their geographical distribution, their transport capacity, or how they might respond to increased meltwater production on the ice surface. Logistical challenges render field research difficult, but advances in remote sensing enhance the ability to monitor dynamics across space and time. Accelerated climate warming increases suitable areas for supraglacial river formation and underscores the importance of coupling supraglacial river hydrology with ice dynamics to improve estimates and projections of eustatic sea level rise.

SUMMARY POINTS

1. Supraglacial streams and rivers are fluvial features that signify active meltwater production and transport on the ablating surfaces of glaciers and ice sheets, while their termination points influence subglacial conditions.
2. Polar explorers documented the presence of supraglacial streams and rivers, and these features received early study during the early to mid-nineteenth century in the context of cold regions' military operations. Scientific investigations now span glaciers and ice shelves in both the Northern and the Southern Hemispheres as well as alpine glaciers.
3. Supraglacial streams and rivers link surface climatology with ice dynamics and thus are an important component of how glaciers respond to climate variability; control surface melt by modulating surface reflectance (albedo) and topographic roughness; and impact the thermal regime of ice by descending into a glacier, freezing, and releasing latent heat.
4. Supraglacial stream and river hydrographs exhibit strong seasonal and diurnal variability. This is driven by surface climatology, catchment geometry, channel density, and hydrologic connectivity in the near-surface weathering crust.
5. Commonalities exist between supraglacial and terrestrial rivers, including drainage pattern, fluvial morphometry, channel incision, and meandering.

FUTURE ISSUES

1. The global distribution of supraglacial streams and rivers remains unknown.
2. Remote sensing and in situ monitoring of supraglacial streams and rivers can help validate and interrogate climate-based meltwater runoff models to better constrain projections of future sea level rise.
3. Ice dynamics models are not yet coupled with surface hydrology routing models, impeding understanding of interactions between climate, surface mass balance, and ice flow.
4. Supraglacial stream and river channel incision models could be tested using field observations to help determine optimal applications for each modeling framework.
5. Understanding of the effect of ice crystal transport on supraglacial stream and river fluvial geomorphology and hydraulics remains preliminary, and further investigations are needed.

6. The production, transport, and export of meltwater on ice shelves and related impacts on ice shelf stability remain poorly understood and are important for understanding the vulnerability of the Greenland and Antarctic Ice Sheets to a warming climate.

DISCLOSURE STATEMENT

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