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Floodplain dynamics in North American permafrost regions under a warming climate and implications for organic carbon stocks: A review and synthesis

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A B S T R A C T

Although there have been studies on changes to hydrology in permafrost regions and exports of nutrients and organic matter to the Arctic Ocean, little is known about how geomorphic dynamics of rivers in permafrost regions will change in the future under a warming climate and the effects of those changes on floodplains. We focus on river dynamics in the context of channel-floodplain interactions and the implications for organic carbon storage in floodplains. As sites of nutrient processing and storage of sediments and organic matter, changes in channel and floodplain form and process will impact sediment yields, nutrient and organic matter export to the Arctic Ocean, aquatic and riparian habitat, and infrastructure. We present a review of the factors influencing reach-scale river dynamics, using the framework of factors affecting erosive force and erosional resistance of banks and floodplain surfaces, which will change due to a warming climate. We summarize studies indicating how observed and modeled trends in these factors will affect erosive force and erosional resistance in the future. We then hypothesize the net effects that these changes will have on the ratio of erosive force to erosional resistance, and the cascading effects on channel and floodplain form and process. We describe two scenarios that could occur under different conditions in the form of conceptual models, one in which the ratio of erosive force and erosional resistance decreases, and one in which the ratio increases. An increase in the ratio of erosive force to erosional resistance due to a reduction in permafrost extent and depth and an overall increase in discharge would increase bank erosion, bank failures, sediment supply, and lateral channel migration rates, decreasing floodplain turnover time and the age of riparian vegetation. A decrease in the ratio of erosive force to erosional resistance due to a reduction in erosive force relative to sediment supply would cause enhanced deposition within the river corridor. Regardless of which scenario may occur, changes in channel process and form will influence the ratio of lateral to vertical accretion, change the nature and stored amount of floodplain sediment, and change the sources and storage of organic carbon within floodplains.

1. Introduction

1.1. Permafrost degradation under a warming climate

The Arctic and Subarctic, which on the North American mainland extend from ~55 to 72°N, are warming faster than other regions in the world (IPCC, 2014; U.S. Environmental Protection Agency, 2016). Warming climate will cause diverse changes in factors that influence channel and floodplain process and form, such as changes in precipitation regimes (Rawlins et al., 2010). One of the most important effects of warming climate at high latitudes is the degradation and loss of permafrost (Fig. 1) (ACIA, 2005; IPCC, 2014; Jorgenson et al., 2006), which is predicted to be severe during the 21st century (IPCC, 2014; Lawrence and Slater, 2005). Permafrost is lithospheric material that has a temperature at or below 0 °C for at least two successive years (Wright et al., 2009). Based on its areal extent, permafrost can be designated as continuous (> 90% coverage), discontinuous (50%–90% coverage), or sporadic (10%–50% coverage). The Arctic is sometimes defined by the limit of continuous permafrost and the Subarctic by the limit of discontinuous permafrost (Woo and Winter, 1993). Permafrost underlies approximately 24% of the land in the northern hemisphere (Zhang et al., 2003) and contains substantial quantities of organic carbon and water in the form of ice (Hugelius et al., 2014). Permafrost is overlain by a seasonally thawing active layer that can vary from a few centimeters to > 1 m in thickness. Near-surface permafrost within 1 m of the surface, which is particularly susceptible to a warming climate, underlies approximately 46% of the Alaskan Yukon River basin (Pastick et al., 2014), highlighting the potential vulnerability of major rivers basins to changes associated with permafrost degradation. The extent of near surface permafrost will likely decrease by the year 2100 (IPCC, 2014; Pastick et al., 2015).

Thawing permafrost will cause many geomorphic changes across the landscape (Rowland et al., 2010). Permafrost degradation and climatic changes are causing, and will continue to cause, changes in: hillslope erosion and thermokarst development and resulting sediment yield to rivers (Gooseff et al., 2009; Lamoureux and Lafrenière, 2009);

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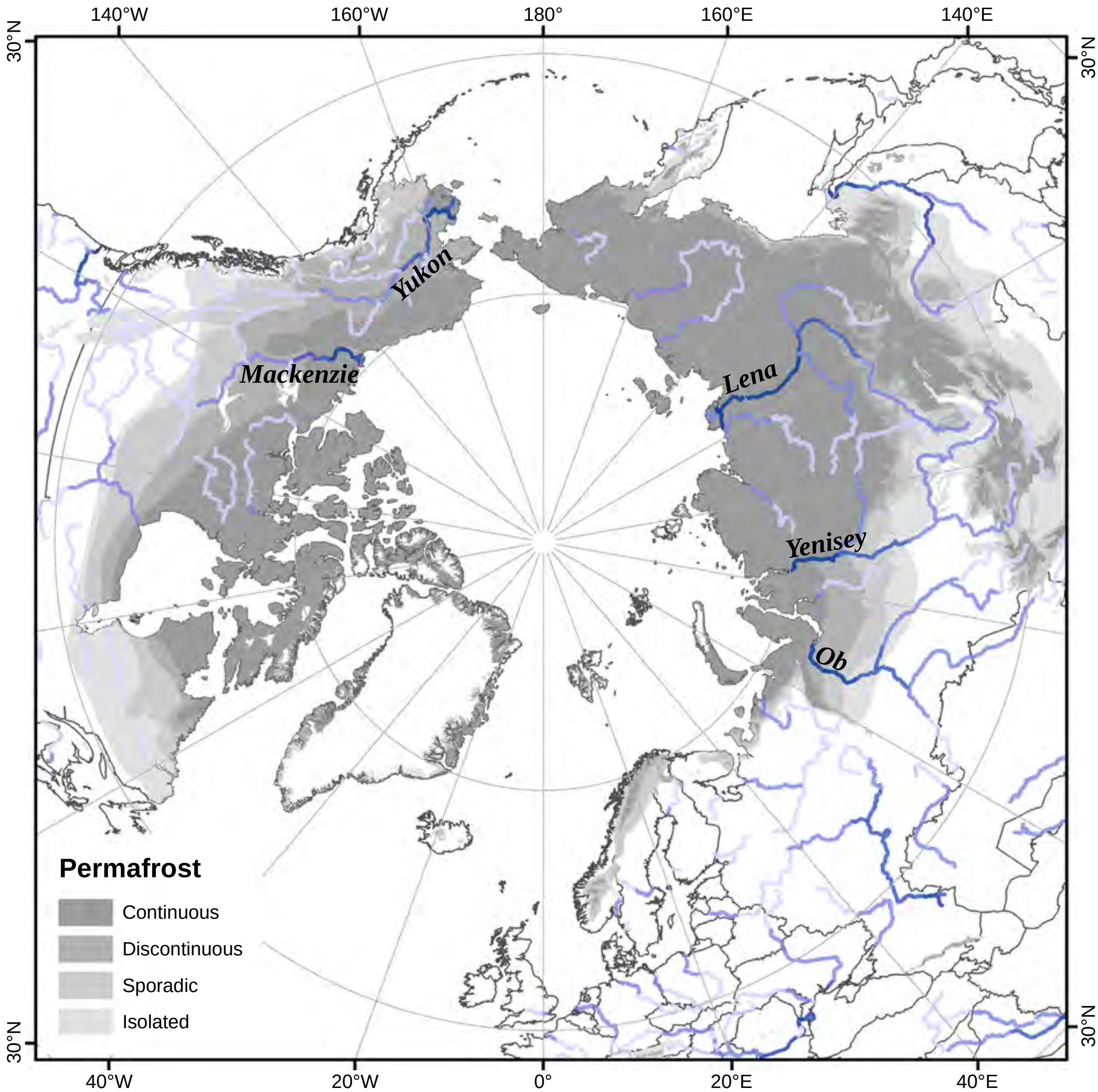

Fig. 1. Map showing the extents of continuous (> 90%), discontinuous (50–90%), sporadic (10–50%), and isolated (< 10%) permafrost in the Arctic and Subarctic (from Brown et al., 2002). Major rivers of the region are also shown, with relative sizes indicated by shading. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

vegetation communities and distributions within watersheds (Danby and Hik, 2007; Osterkamp et al., 2009); runoff, sediment, and organic matter delivered to streams and rivers (Walvoord and Kurylyk, 2016; Walvoord and Striegl, 2007); river hydrology (Overeem and Syvitski, 2010; Woo et al., 2008b) and sediment transport (Dugan et al., 2009; Kokelj et al., 2013); nutrient delivery to the oceans via rivers (Frey and McClelland, 2009; Guo et al., 2007); and organic carbon oxidation, erosion, and riverine transport (Schuur et al., 2015; Vonk and Gustafsson, 2013).

1.2. River dynamics as a mediator of permafrost–carbon fluxes

Geomorphic changes to the landscape caused by permafrost degradation will impact sediment and carbon dynamics within channels and floodplains. Soil carbon is the largest terrestrial organic carbon stock, and permafrost regions contain an estimated 1300 Pg (1 Petagram = 10^{15} g) of carbon in the subsurface (Hugelius et al., 2014; Jobbágy and Jackson, 2000). Thawing permafrost could release large amounts of currently frozen carbon into the atmosphere through microbial decomposition and outgassing (Schuur et al., 2015; Schuur et al., 2008), and these releases could cause a positive carbon feedback and further warming of global temperatures (Koven et al., 2011; Schuur et al., 2015). In addition, carbon from previously frozen ground can enter surface runoff, groundwater, and rivers, becoming available for microbial respiration and/or storage in floodplains and deltas before being deposited in the ocean (Hilton et al., 2015; Vonk and Gustafsson, 2013). Thus, it is of vital importance to understand the details of how carbon released from thawed ground is likely to move through freshwater networks along with sediment, as well as the geomorphic changes to potential transient storage areas for sediment and carbon, such as floodplains.

Despite numerous papers addressing the likely effects of warming climate and permafrost thaw on hydrologic cycles (Walvoord and Striegl, 2007) and riverine transport of dissolved and particulate matter in permafrost regions (Frey and McClelland, 2009; Hilton et al., 2015; Raymond et al., 2007; Syvitski, 2002), little work has been done on how channel and floodplain geomorphic dynamics might change (but see Ashmore and Church (2001)). The great rivers of the high latitudes in the northern hemisphere – the Yukon, Ob-Irtysh, Lena, and others – have extensive floodplains underlain by permafrost. Estimates based on Digital Elevation Models (1 km resolution), drainage networks, and a flooding algorithm suggest that deltas and floodplains comprise ~7 ± 4% of the land area in circumpolar permafrost regions (Shelef, 2019, personal communication, February 21, 2019). These permafrost-rich floodplains are topped by an active layer that also contains substantial organic carbon stocks in part because of the extensive wetlands facilitated by the underlying permafrost and associated limited infiltration capacity (Lininger et al., 2019). Warming-induced changes in hydrologic processes, sediment movement, and vegetation will result in altered riverine fluxes to the ocean, but these altered fluxes will be mediated by changes in channel-floodplain interactions.

We define a floodplain as the alluvial landform adjacent to a channel, separated from the channel by banks, and built of sediment that has been primarily transported by the present flow regime, although past conditions may have created older floodplain surfaces (Nanson and Croke, 1992). We focus on changing floodplain dynamics, but floodplains are created as a result of channel processes, and thus much of this review discusses relevant channel processes. Floodplains are built by lateral and vertical accretion of sediment carried by rivers and reflect the balance between a river's ability to erode its floodplain and transport incoming sediment (erosive force) and the resistance of the banks and floodplain surface to erosion or mass bank failure (erosional resistance) (Nanson and Croke, 1992; Simon et al., 2000; Wolman and Leopold, 1957). Although bank and bed process and form are inter-related, we focus on banks as the interface between channels and floodplains and consider them as boundaries of the floodplain surface.

Floodplains are important regions of nutrient and water exchanges (Appling et al., 2014; Junk et al., 1989; Lininger et al., 2018; Mertes, 1997) and contain significant organic carbon (OC) stocks in sediment and downed wood (Cierjacks et al., 2010; Lininger et al., 2019; Lininger et al., 2017; Sutfin et al., 2016; Wohl et al., 2012). Floodplains also reduce flood peaks thr ough storage of water associated with attenuation and storage of flood waves (Lininger and Latrubesse, 2016; Sholtes and Doyle, 2011). Floodplains, particularly on large rivers, can store sediment within a river network for long periods of time (10^2–10^3 years), and a significant portion of sediment delivered to river outlets can be sourced from the floodplain as opposed to headwaters (Aalto et al., 2008; Dunne et al., 1998; Mertes et al., 1996). Floodplains provide habitat for aquatic and terrestrial species and can increase biodiversity by increasing spatial diversity of habitats (Bellmore et al., 2012; Ward et al., 2002). As floodplain dynamics change due to permafrost thaw and altered bank erosion, there will likely be changes in sediment dynamics and nutrient cycling within river systems (Frey and McClelland, 2009; Lewis and Lamoureux, 2010), resulting in changed exports to estuaries and near-shore ocean environments (Raymond et al., 2007; Syvitski, 2002). In addition, there will likely be substantial negative impacts on infrastructure within floodplains as permafrost thaws.

1.3. River dynamics during past climatic changes

Studies of past warming episodes in regions of permafrost have provided insight into potential changes that may occur due to current warming. As drainage basins in southern England underlain by permafrost during the Pleistocene transitioned to a warmer Holocene climate with more extensive woody vegetation, they went through a phase of greater sediment production that increased lateral channel movement and floodplain aggradation (Murton and Belshaw, 2011). Within permafrost regions, warming during the Pleistocene-Holocene transition corresponded to increased sediment inputs, expansion of woodlands, peatlands, and thermokarst, and floodplain alluviation (Mann et al., 2010). Mann et al. (2010) describe rapid floodplain alluviation and aggradation on rivers along the northern slope of Alaska's Brooks Range during this warming period. They attribute this change to excess sediment supply resulting from thermokarst and mass failures on hillslopes. In addition, records of channel response from different regions suggest that changes in river dynamics due to climate may be mediated by variations in soil cohesion and peak discharges, which are related to vegetation and permafrost characteristics (Vandenberghe, 2008; Vandenberghe, 2002). These studies of past climatic changes indicate that dramatic changes are likely to occur with climate warming, and there is a need for a generalized framework applicable to many river systems that will be testable as changes occur.

1.4. Objectives and structure of this paper

Our primary objective in this paper is to build on contemporary understanding of (i) channel and floodplain processes in North American rivers underlain by permafrost and (ii) predicted changes in high latitude climate and hydrology in North America to develop conceptual models of changes in river and floodplain dynamics and the resulting changes in OC dynamics. We do not intend these conceptual models to be site-specific predictions, but rather syntheses of potential scenarios of change based on past climatic alterations, observations of response to ongoing climate warming, and extrapolations based on knowledge of floodplain and organic carbon dynamics in permafrost regions.

Many of the processes that we discuss and the inferences that we make are also applicable to rivers in permafrost regions of Eurasia. We focus on North American rivers because we have direct experience with the diversity of rivers in North American permafrost regions. Within the North American permafrost region, the Yukon and Mackenzie are the largest rivers, covering 854,700 and 1,810,000 km^2, respectively. Each of these enormous river drainages includes substantial diversity of river process and form, and we focus on these drainage basins, while also drawing on the literature describing river dynamics in regions such as the Brooks Range and coastal plain on the North Slope of Alaska and the low-relief terrain of far northeastern Canada.

We first discuss the distinctive characteristics of rivers in northern permafrost regions, including ice within channels and the presence of permafrost. Working within the framework of understanding river dynamics as erosive driving forces (water, sediment, ice) versus erosional resistance of the floodplain (including floodplain surfaces and channel banks), we review the main factors influencing erosive force and erosional resistance, focusing on factors that will likely change with increasing temperatures in regions with permafrost (Fig. 1). Next, we review the changes in these main factors that are occurring and will occur in the future due to global warming and examine the potential effects on erosive force and erosional resistance. We present two scenarios of how warming will influence channel and floodplain process and form through changes to the ratio of erosive force to erosional resistance, and then describe regional variations that influence the effects of these changes. Finally, we discuss the implications of changes in river dynamics, highlighting impacts to sediment yields, OC cycling, aquatic and riparian habitat, and infrastructure.

2. Distinctive characteristics of rivers in northern permafrost regions

Rivers in regions underlain by permafrost exhibit diverse morphology and processes (Vandenberghe and Woo, 2002), but they commonly have distinctive characteristics relative to rivers at lower latitudes because of the presence of permafrost and seasonal ice cover on the rivers. Although there may also be variations in precipitation regimes, vegetation, and other characteristics within river basins in permafrost regions, we focus on the presence of permafrost and river ice because these distinctions are likely the most different compared to regions that are permafrost-free. Changes in the characteristics of both permafrost and seasonal ice cover as climate warms will likely create complex, nonlinear interactions among precipitation and runoff to streams, bank stability, river erosion, and channel and floodplain morphology (Phillips, 2003). In this section, we review general hydroclimatic patterns in North American permafrost regions, the effects of permafrost and seasonal ice cover on river process and form, and organic carbon dynamics of rivers in permafrost regions.

2.1. Hydroclimatic patterns of North American permafrost regions

North American permafrost regions are primarily characterized by short, moderately warm, and moist summers and long, extremely cold, and dry winters (Bonan and Shugart, 1989). Except for the western coastal fringe, mean annual precipitation in the North American Arctic and Subarctic is mostly < 800 mm. Arctic precipitation contains a mixture of water originating from evaporation of local sources and remote sources: the ratio between these local and remote sources fluctuates seasonally (Bintanja and Selten, 2014).

Hydrologic regimes in permafrost regions are varied depending on sources and runoff characteristics. Although different classifications exist, hydrograph regimes can be generally classified as nival (Arctic or Subarctic), in which large flows are generated by snowmelt and river ice break-up; proglacial, in which high flows last into summer as a result of glacial melt; wetland, in which rivers draining wetlands have prominent snowmelt peaks because the wetlands have low storage capacity while frozen, but can retain water and retard summer flows later in the warm season; and prolacustrine, in which rivers below large lakes have a fairly even runoff throughout the year (Prowse, 1994; Woo and Thorne, 2003). In this paper, we focus on rivers with nival and wetland regimes because proglacial and prolacustrine regimes have distinctive hydrologic and geomorphic characteristics and would require a more extensive review. However, some of the concepts reviewed in this paper could also apply to these regimes, such as the factors influencing erosive force and erosional resistance.

2.2. The influence of permafrost on hydrology, sediment loads, and bank resistance

One of the primary distinctive characteristics of rivers in permafrost regions is that the impacts of precipitation on spatial and temporal variations in runoff to rivers are uniquely mediated by permafrost extent and thawing in the active layer (Wang et al., 2009; Woo et al., 2008a). Permafrost limits infiltration from rainfall, snowmelt, and melting ice within the active layer, which occurs during warmer months. This results in greater water yield and larger direct runoff ratios for rain and snowmelt relative to non-frozen or temperate regions (Hayashi et al., 2003; Woo et al., 2008a). Most rivers in permafrost regions experience highest flows during snowmelt and tend to have little to no baseflow during winter, when frozen conditions limit the delivery of water into river networks (Woo et al., 2008a).

Permafrost and variations in active layer characteristics influence the spatial patterns and proportions of overland and subsurface flow. Subsurface flow in areas of permafrost can occur above or below permafrost and within unfrozen zones (taliks) (Walvoord and Kurylyk, 2016). Permafrost can support a high riparian water table that leads to substantial overland flow during precipitation or snowmelt and limits inputs of upland groundwater to channels (Dingman, 1973). This results in a flashy hydrologic response relative to basins without permafrost (Woo, 1986). In contrast, overland flow is not common on hillslopes that contain permeable moss and peat (Dingman, 1973; Wright et al., 2009). On this type of slope, runoff occurs primarily as subsurface flow in the seasonally thawed, saturated layer above the permafrost. The depth of this layer changes during the thaw season (Woo, 1986), creating temporal changes in surface infiltration and soil water-retention capacity. The direct runoff ratio and the rate of hydrograph recession can decrease with time during the melt season as the thickness of the thawed portion of the active layer progressively increases (Yamazaki et al., 2005). In areas with discontinuous permafrost, spatially variable storage within the active layer dictates the location and extent of subsurface flow because thawed areas have to be filled before water can spill and generate downslope flow (Wright et al., 2009). As Wright et al. (2009) demonstrate, the feedbacks between thawing and subsurface flow in discontinuous permafrost regions are particularly complex because of spatial and temporal variability in active layer depths and the heat transfer involving water.

The presence of permafrost also supports extensive floodplain wetlands even in regions where precipitation totals are relatively low; more than half of all wetlands are located in permafrost-influenced regions of northern high latitudes (Avis et al., 2011). Because of the confining permafrost layer, subsurface flow in wetlands is restricted to a shallow layer where permafrost is present, but in discontinuous permafrost wetlands may interact with deeper groundwater (Woo and Winter, 1993).

Relative to other climatic regions, Arctic and Subarctic rivers discharge less sediment to the ocean, although there are variations in this general trend (Syvitski, 2002). This is likely due to many factors, including the presence of frozen ground and the high proportion of lakes in high latitude regions, which can filter out sediment (Syvitski, 2002). However, the Yukon and Mackenzie contribute about one-fifth of the yearly water discharge of the world's eight largest Arctic rivers, but contribute almost three-fourths of the suspended sediments, likely because they drain tectonically-active and glaciated landscapes relative to other large Arctic rivers (Holmes et al., 2002).

The presence of permafrost can influence sediment load within high latitude rivers due to thermokarst development, in which ice-rich permafrost thaws and causes surface depressions in the landscape (Olefeldt et al., 2016). Mega slumps, which can cover up to 40 ha and develop headwalls up to 25 m in height, are associated with debris flows and distinct patterns of river discharge, solute flux, and sediment concentrations across a range of watershed scales (Kokelj et al., 2013). Many of the papers describing thermokarst focus on field areas on the North Slope of Alaska, where thick alluvial deposits facilitate extensive thermokarst (Farquharson et al., 2016; Gooseff et al., 2009), but thermokarst landscapes occur across northern North America (Olefeldt et al., 2016).

Permafrost also strongly influences the erosional resistance of banks and the processes by which banks are eroded (Costard et al., 2003; Scott, 1978). Thermal erosion, in particular, is a distinctive aspect of river dynamics in permafrost regions. During thermal erosion, heat transfer from the flow of water through the frozen ground thaws the ground, allowing banks to collapse or be abraded more easily (Costard et al., 2003). Higher water and ice temperature, as well as increased discharge, accelerate thermal erosion, whereas greater ice content in the bank sediment retards thermal erosion (Randriamazaoro et al., 2007). Laboratory experiments examining the effect of bank sediment texture and ice content indicate that greater ice content can either increase or decrease thermal erosion, depending on the ice content and the interactions among ice content and spatial distribution, bank sediment texture, and stream flow velocity and turbulence (Dupeyrat et al., 2011). Of the parameters influencing thermal erosion, the effects of water temperature and high discharges are particularly important (Costard et al., 2003; Randriamazaoro et al., 2007). Thermal erosion occurs predominantly during a few weeks in spring when discharge is high and more of the bank is in contact with water. Randriamazaoro et al. (2007) describe an initial acceleration phase with an increasing thermal erosion rate that commonly lasts for a few days when the water temperature is close to the melting point.

The greatest rates of thermal bank erosion occur in non-cohesive sediments, particularly at concave river banks and the head of islands (Costard et al., 2003). If detached bank sediment accumulates at the base of the eroding bank, as can occur along very gently sloped banks or in low-velocity portions of the channel margins, the accumulated sediment can limit or prevent further erosion (Costard et al., 2003). Thermal erosion is thus not spatially or temporally uniform, but instead varies with factors such as timing within the hydrograph (water temperature, discharge), bank composition, bank height and angle, and bank exposure to hydraulic forces (Costard et al., 2003).

In summary, the effects of permafrost on infiltration, runoff, riparian water table, sediment loading, and erosional resistance of banks are spatially and temporally variable. These effects are characterized by nonlinear interactions with processes such as precipitation inputs and river flow.

2.3. Hydrologic and geomorphic effects of river ice

A second distinctive characteristic of rivers in permafrost regions is the presence of seasonal ice cover on the active channel. Just as the presence of permafrost and an active layer can complicate rainfall-runoff relations and river stage, the presence of ice cover can complicate river stage-discharge relations. A river-ice season of > 100 days duration between autumn freeze-over and spring breakup characterizes many rivers of permafrost regions (Prowse and Beltaos, 2002).

River ice can directly modify the quantity of river flow via: (1) anchor ice frozen to the bed of the channel that cuts off groundwater inflow; (2) direct storage of water in river ice, which commonly involves slower removal during freeze-over and rapid resupply during breakup; and (3) hydraulic storage in the channel when hydraulic resistance associated with ice cover increases river stage, especially when the ice cover is hydraulically roughest during freeze-over and breakup and when ice jams form during breakup (Ettema and Kempema, 2012; Prowse and Beltaos, 2002). Reduction in river flow created by these three mechanisms can be equivalent to nearly 30% of the flow that would otherwise occur, and release of stored water during breakup can account for nearly 20% of the spring peak flow (Prowse and Carter, 2002). Ice effects can result in the lowest discharge of the year during winter freeze-over due to storage of water in ice, even though runoff is lowest during late winter. Annual maximum river levels commonly occur during ice breakup, even though maximum discharge is more likely to result from spring snowmelt or summer rainfall later in the season (Prowse and Ferrick, 2002).

Icings or aufeis form a distinctive subset of river ice. These sheet-like masses of layered ice form during the winter via freezing of successive flows of water on top of river ice (Morse and Wolfe, 2015). The water that creates icings originates from groundwater springs, which can be important components of water storage in unglaciated basins (Morse and Wolfe, 2015). Icings are spatially recurrent and can be > 10 km^2 in area and > 10 m thick (Pavelsky and Zarnetske, 2017), but do not necessarily occur every year or cover the same spatial extent. Like other forms of river ice, icings create flow barriers during the spring freshet and increase flooding potential.

Break-up of river ice can be thermal or mechanical events (Beltaos and Prowse, 2009). Thermal events occur when river ice has been weakened due to warming, and these events cause less flooding along with ice break-up. In contrast, mechanical events occur when thick, strong ice is broken by strong hydraulic force and can cause ice jam floods and greater geomorphic impacts (Beltaos and Prowse, 2009).

River ice moving downstream can also influence channel morphology through mechanical erosion of banks and alteration of bank vegetation. An extensive literature documents freeze-thaw cycles as an effective mechanism for weakening the cohesion and erosional resistance of banks and for detaching individual sediment grains or aggregates of sediment (Lawler, 1992; Leopold, 1973; Pizzuto, 2009; Wolman, 1959; Wynn et al., 2008; Yumoto et al., 2006). Ice abrading the banks while in transport can also be very effective. The effectiveness of this abrasion can vary in a nonlinear manner during a single peak flow season. Costard et al. (2014) describe accumulation of ice along river banks and islands during the initial stage of breakup. These accumulations protect the underlying feature from abrasion for a few days, followed by a period of highly effective erosion after the ice has melted. Smith (1979) describes channel enlargement associated with scour during ice breakup and transport. When ice jams recur with sufficient frequency, ice scour can produce a two-level bank morphology with a much wider upper channel at the level of ice scour (Boucher et al., 2012, 2009). By severely abrading or destroying bank vegetation, river ice jams can reset the vegetation succession to the initial point (Uunila and Church, 2015), thus limiting the above- and below-ground effects of vegetation on hydraulic force and bank erosional resistance.

Ice jams within the channel during breakup can reroute flow across floodplains and cause channel avulsion (Burge and Lapointe, 2005; Gay et al., 1998; Hicks, 1993) or headward gully erosion where flow returns to the channel. Smith and Pearce (2002) infer this process to be a precursor to the formation of the oxbow lakes and multiple channels that are common in many northern rivers. Ice blocks can also be carried onto floodplains and grounded, facilitating the formation of water vortices beneath and between the blocks and erosion of scour holes by the vortices (Smith and Pearce, 2002). The Mackenzie River at Fort Providence may experience avulsions as a result of ice jams (Hicks, 1993).

In summary, the presence of seasonal ice cover can strongly influence river stage during various seasons of the year. Breakup of the ice cover and ice abrasion can create substantial erosive forces directed against the banks and floodplain surface.

2.4. Organic carbon dynamics in North American rivers with permafrost

Rivers in permafrost regions transport large amounts of dissolved organic carbon (DOC) relative to rivers in lower latitudes. Comparisons of watersheds in discontinuous permafrost regions indicate that those with higher percentages of permafrost have higher concentrations of DOC of terrestrial origin within the river network (Balcarczyk et al., 2009). Fluxes of DOC in rivers with permafrost predominantly occur within a relatively short period after spring ice break-up and are much greater than fluxes from equivalently sized rivers of temperate latitudes (Raymond et al., 2007). Relative to DOC, large rivers in the high latitudes discharge less particulate organic carbon (POC) to the Arctic Ocean (McClelland et al., 2016). Although understudied, floodplains in permafrost regions may store more OC in floodplain soils and downed large wood relative to upland environments (Lininger et al., 2019, 2017).

The age of POC being transported to the Arctic Ocean, which can be tens of thousands of years old (Hilton et al., 2015; Schreiner et al., 2014), suggests that POC may be sourced from river banks containing old permafrost. DOC, although it can be old when first released from permafrost, is rapidly lost through microbial respiration in both upland (Zimov et al., 2006) and riverine (Spencer et al., 2015) environments, resulting in relatively young DOC exported to the Arctic Ocean (Aiken et al., 2014; Raymond et al., 2007). Modern DOC may also be flushed directly from surface vegetation and peatlands into high latitude rivers and exported to the Arctic Ocean (Raymond et al., 2007).

3. Erosive forces and erosional resistance of banks and floodplain surfaces

Erosive forces exerted against the channel banks and floodplain surface in association with river flow include shear stress, thermal erosion, and abrasion by ice blocks. Shear stress and thermal erosion increase with increasing discharge. Ice abrasion, as explained in section 2.3, can increase with increasing discharge, but does not necessarily do so, especially during the early stages of breakup or ice-jam floods.

Erosional resistance of banks, i.e., shear strength, results from the frictional properties of sediments, the effective normal stress of the bank, and effective cohesion (Simon et al., 2000). Effective normal stress is the normal stress of a bank, which acts to hold the bank in place, minus the pore water pressure, which acts to push apart sediment grains. Effective cohesion results from cohesion added from the roots of riparian vegetation, matric suction in unsaturated sediment, and cohesive forces (Eaton, 2006), including those associated with the presence of permafrost. Bank erosion occurs via detachment of individual grains or via mass failure. Erosional resistance of floodplain surfaces results from the frictional properties of the sediment and from effective cohesion, but it is also influenced by the flow resistance created by living vegetation (Järvelä, 2004), downed, dead wood (Wohl, 2013), and local floodplain topographic relief (Güneralp and Rhoads, 2011). As with channel banks, the roots of floodplain vegetation can also increase erosional resistance of the floodplain surface.

In this section we review the factors that influence the erosive forces exerted against banks and floodplain surfaces and the erosional resistance of banks and floodplains. We focus on factors that will likely be influenced by climate change.

3.1. Factors influenced by climate change

3.1.1. Basin-scale factors

Although not the focus of this paper, we recognize the importance of basin-scale factors that influence floodplain process and form, such as geologic controls associated with tectonics and relative base level, and climatic controls such as sea level fluctuations (Dunne and Aalto, 2013). Some of these basin-scale factors will change under a warming climate. Base-level changes (i.e., sea level fluctuations) can influence aggradation and degradation within a drainage basin (Schumm, 1993). For example, sea-level rise during the Holocene promoted large-scale sediment storage in the lower valley of the Amazon River (Mertes and Dunne, 2007). Although these large-scale, long-term processes will affect rivers underlain by permafrost, we focus on local-scale factors influencing erosive force and erosional resistance that will change with a warming climate, likely at shorter time scales.

3.1.2. Factors influencing erosive force

We identify six main factors influencing the erosive force exerted against floodplains and floodplain boundaries (banks) in permafrost regions that are currently changing or will change in the future because of global warming: magnitude of discharge and duration of peak discharge ($Q_{mag,dur}$), water temperature, sediment load (Q_s), dispersed large wood (LW) load (volume of LW per area of channel), the number and load of LW jams, and frequency and extent of ice jams (IJ). Table 1 displays the effect of each factor on erosive force. These responses are meant to be broadly applicable to diverse river geometries. We recognize that multiple factors will be changing simultaneously and interacting with each other, but we start by examining relations among pairs of variables. If other factors remain steady or consistent, for example, increasing water discharge creates greater erosive force against the floodplain boundaries. In addition to magnitude, a longer duration of high magnitude discharges increases the duration of significant force that is applied to floodplain boundaries. The magnitude and duration of peak discharge also influence overbank flooding, and the potential for stripping of the floodplain surface.

Water temperature strongly influences thermal erosion of channel banks with permafrost (Costard et al., 2003). We draw on the work of Randriamazaoro et al. (2007) to hypothesize an increase in erosive force with increasing water temperature, both during seasonal peak flow, other flows of the hydrograph, and over a period of years.

Sediment load can have variable impacts on erosive force. Increasing discharge of bed-material load (grain sizes found in appreciable quantities on the bed) can cause increased deposition on point bars in the inner bank of meander bends, deflecting flow towards the outer banks and increasing bank erosion (Dunne et al., 2010). Meandering rivers with greater sediment supply have been shown to have faster channel migration rates and more frequent meander cutoffs from increased bank erosion (Constantine et al., 2014), indicating greater erosive force exerted against the floodplain boundary. Increasing bed-material fluxes can also be associated with braiding and enhanced bank erosion (Ashmore, 2013). Wash load concentrations (silts and clays traveling only in suspension) are relatively low in northern rivers (Syvitski, 2002). Thus, although wash load requires energy for sediment transport, increases viscosity, and may dampen turbulent fluctuations that can result in shearing force directed at the channel boundaries at high concentrations (Kuhnle, 2013), the influence of wash load on erosive force in permafrost rivers will likely be negligible. We recognize that changes in wash load can have indirect effects on bank erosion: for example, increasing suspended sediment concentration can enhance transport of bedload (Simons et al., 1963).

Sediment load can also influence erosive force by overwhelming the capacity of the river flow to transport sediment, causing aggradation and effectively reducing erosive force relative to sediment load (Leopold et al., 1964). Increases in sediment supply, for example due to past thermokarst activity during Pleistocene-Holocene transitional warming periods, can cause increased sedimentation within valleys (Mann et al., 2010).

Dispersed pieces of large wood within the bankfull channel can increase hydraulic roughness and thus reduce velocity and erosive force exerted against the banks (Brooks et al., 2003; Manga and Kirchner, 2000). Large wood that falls into a channel while remaining attached to

Table 1
Factors influencing erosive force, showing the directional change in erosive force with an increase in the factor.

Factor	Change in erosive force with increase in factor
Discharge (Duration, Magnitude)	+
Water temperature	+
Sediment load	±
Dispersed Large Wood	
Stationary	-
In transport	+
Large Wood Jams	
Upstream from the jam	-
At	±
Downstream	±
Ice Jams	
Upstream from the jam	-
At	±
Downstream	+

Fig. 2. Examples from the Yukon River drainage basin of interior Alaska of large wood falling from banks and reducing hydraulic force exerted against banks.

the top of the bank via the roots (Fig. 2) can be particularly effective at increasing hydraulic roughness along the bank and reducing erosive force (Daniels and Rhoads, 2004).

Concentrations of large wood in the form of jams can have varied localized effects on erosive force upstream, at, or downstream of the jam, and the effects depend on whether the jam is mobile or stable. A jam can (i) create a backwater that substantially reduces velocity and bank erosion upstream (Le Lay et al., 2013; Triska, 1984), (ii) deflect flow towards the bank, promote bar growth and lateral channel movement (Collins et al., 2012; Martín-Vide et al., 2014; O'Connor et al., 2003), or move as congested transport (Braudrick et al., 1997) with the potential for substantial bank erosion at the jam or downstream of the jam (Piegay, 1993), or (iii) remain stable over longer time periods and cause sufficient flow obstruction to promote overbank flow and dissipate flow energy (Triska, 1984). The influence of jams on geomorphic processes can depend, however, on whether the jams are channel-spanning or located on channel margins. Channel-spanning jams are more likely to deflect flow overbank or create backwater effects (Jeffries et al., 2003). LW in transport can physically abrade floodplain surfaces and living vegetation (Johnson et al., 2000). LW can direct and concentrate overbank flow across specific portions of the floodplain and initiate splays or avulsion channels (Collins et al., 2012).

The frequency of ice jams over multiple years and the spatial extent of these jams can substantially influence hydrology and hydraulics on rivers with continuous ice cover (Hicks and Beltaos, 2008; Prowse and Ferrick, 2002). Ice jams develop when ice cover breaks up and discharge rises in the spring, and less commonly develop during freeze-up (Hicks, 2009). Although other river ice processes, such as the formation of frazil ice (granular ice transported within rivers) or anchor ice (ice attached to a channel bed) (Hicks, 2009) impact erosive force to some degree, we focus on ice jams as the most important way in which ice exerts erosive force against channel banks. Similarly, resistance added by a stable ice cover elevates water levels in a channel, particularly when the ice cover is hydraulically most rough during freeze-over and break-up (Prowse and Beltaos, 2002), but ice jams likely have the most important influence on erosive force.

Ice jamming can increase peak water levels and cause accelerated bank erosion by gouging the banks, at times promoting meander cutoffs and avulsions (Burge and Lapointe, 2005; Ettema, 2006; Ettema and Daly, 2004; Ettema and Kempema, 2012; Turcotte et al., 2011). Ice jams can direct and concentrate overbank flow and cause erosion into the floodplain surface due to scour around ice blocks (Smith and Pearce, 2002), but ice can also protect banks and reduce the erosive force exerted again them (Costard et al., 2014). Ice jams can also create backwaters, reducing velocities and upstream force upstream (Hicks, 2009; Prowse and Culp, 2003). The surge resulting from break up of an ice jam can enhance overbank flooding and flood wave velocity downstream and this, along with the mechanical effects of large chunks of mobile ice, can enhance bank erosion and force channel avulsion (Beltaos, 2002; Ettema and Kempema, 2012; Prowse and Culp, 2003; Turcotte et al., 2011).

3.1.3. *Factors influencing erosional resistance*

The factors that are most likely to change due to global warming and that influence erosional resistance include the density, rooting depth, and type of riparian vegetation; logjams buried within the floodplain; the lateral extent and depth of permafrost; the characteristics of flood hydrographs; the frequency of freeze/thaw cycling; and grain size of banks. Table 2 summarizes the influence of these factors on erosional resistance.

Riparian vegetation has diverse and sometimes contrasting effects on bank resistance to erosion, but the net effect is likely to be greater bank resistance with increasing spatial density and rooting depth of vegetation (Table 2) (Merritt, 2013). Plants can increase the mass of stream banks, for example, which can promote failure, but this is likely to be counteracted by: (i) increased flow resistance created by the

Table 2
Factors influencing erosional resistance, showing the directional change in erosional resistance with an increase in the factor.

Factor	Change in erosional resistance with increase in factor
Vegetation	+
Buried floodplain logjams	+
Permafrost	+
Discharge (Duration, Magnitude, Variation)	-
Freeze/Thaw	-
Grain size	variable

above-ground portion of the plants, along with reduced near-bank velocity and shear stress (Gorrick and Rodríguez, 2012; Griffin et al., 2005); and (ii) plant roots that increase resistance of bank sediment to shearing (Pollen and Simon, 2005). The type (woody, herbaceous, grasses) and age of vegetation matter as these plant characteristics determine size, stem flexibility, and spatial density of above-ground portions of the plant and the depth, lateral extent, and tensile strength of plant roots (David et al., 2009; Polvi et al., 2014). Small streams with grassy vegetation can be narrower than forested streams because the rooting depth of the densely growing grass roots is similar to that of bank height and the channel migrates laterally relatively quickly but vegetation also regrows quickly (Allmendinger et al., 2005; Hession et al., 2003). In general, woody vegetation results in greater bank resistance compared to non-woody vegetation (Polvi et al., 2014).

One aspect of vegetation that may be unique to large rivers in continuous permafrost zones is the inability of plant roots to penetrate permafrost. Consequently, relatively shallow rooting depths can limit the effectiveness of riparian plants in stabilizing tall banks along large rivers. In this scenario, the collapse of the vegetated active layer once the underlying permafrost is eroded (e.g., via thermal erosion) can create a layer of downed trees resting on or along the bank that may protect the bank from erosive force (see section 3.1.2).

Buried logjams can locally increase floodplain erosional resistance, as described for temperate zone rivers (Collins et al., 2012). Although little information exists on buried logjams in rivers underlain by permafrost, we have observed such jams in the Yukon Flats region of interior Alaska (Fig. 3), indicating that they are present and may influence floodplain erosional resistance.

Permafrost substantially increases bank cohesion, so the lateral extent and depth of permafrost exert a strong control on erosional resistance (Fig. 4) (Costard et al., 2014, 2007; Scott, 1978). When bank permafrost melts via fluvio-thermal erosion or exposure to air, the rate of bank failure and floodplain erosion greatly increases (Costard et al., 2007; Scott, 1978). In addition, melting permafrost could increase pore water pressure, which reduces erosional resistance (Simon et al., 2000).

Discharge magnitude, the duration of peak discharges, and discharge fluctuations during floods also influence erosional resistance. Infiltration into the banks during peak discharges can increase pore water content of the banks (Rinaldi et al., 2004), although this effect may be limited in continuous permafrost regions due to impermeable permafrost within banks. During the falling limb of a flood, the pressure exerted against the banks at the highest stage is removed, promoting bank collapse (Alvarez and Schmeeckle, 2013; Rinaldi et al., 2004). In addition, longer duration floods with multiple peaks (i.e., more complex and varied flood hydrographs) cause greater reduction in erosional resistance because of fluctuations in moisture content within banks and lateral support of banks by water within the channel (Rinaldi et al., 2008, 2004).

Fig. 3. A large buried logjam along the Yukon River that was previously preserved in permafrost. Bank is approximately 5 m high.

Freeze-thaw cycles can substantially decrease erosional resistance by reducing frictional forces between sediment particles and increasing pore water content during melting (Gatto, 1995; Lawler, 1993; Pizzuto, 2009; Wynn et al., 2008; Yumoto et al., 2006). Bank sediments subjected to multiple freeze/thaw cycles can become cohesionless, creating macro-cracks (Kimiaghalam et al., 2015). Consequently, increasing frequency of freeze-thaw cycles is likely to decrease erosional resistance.

Grain size variations are also well documented influences on cohesion of banks and the floodplain surface (Rinaldi and Darby, 2007). Clays and silts are difficult to erode due to their cohesion, while sand requires lower velocities for entrainment; larger grain sizes are difficult to entrain due to their larger mass (Hjulstrom, 1935).

3.1.4. Interactions between factors

We have described the direct influences of individual factors on erosive force and erosional resistance, but most of the factors described above also have indirect effects. For example, the presence and extent of ice cover modifies erosive force by influencing the magnitude of discharge during winter and spring. Spring peak flow magnitude can be substantially increased by release of water stored upstream from ice-covered river segments and by release of water within the ice (Prowse and Carter, 2002), thus increasing erosive force. In addition, water stored in ice during the winter reduces base flows (Prowse and Carter, 2002), decreasing erosive force. Ice can also damage or remove riparian vegetation and reduce the ability of vegetation to grow on banks (Fig. 5) (Ettema and Daly, 2004; Nilsson et al., 2012; Uunila, 1997), which then reduces bank resistance.

Discharge magnitude and duration influence permafrost extent and location because high flows of long duration can cause thermal erosion (Costard et al., 2014). Thus, higher magnitude and longer duration of peak discharges not only increase erosive force but also decrease erosional resistance via a reduction in permafrost extent. Decreased permafrost extent can lead to more large wood delivery to the channel via bank collapse (Ott et al., 2001), resulting in decreased erosive force if the wood is dispersed and various effects on erosive force if the wood is in jams. Bank collapse due to permafrost reduction can also increase the bed-material load and wash load within a reach, with subsequent effects on erosive force.

The spatial distribution of woody vegetation on within-channel bars and islands can indirectly influence bank erosion by forming depositional sites during bedload-transporting flows. The obstructions to flow created by vegetated bars and islands can exacerbate the erosive force exerted against channel banks, enhancing bank erosion and lateral channel migration (McKenney et al., 1995).

Although these indirect effects and interactions among individual factors are likely to substantially influence erosive force and bank resistance, the nature and magnitude of these influences may be more site-specific and thus are difficult to describe in terms of the general trends that are the focus of this paper.

4. Observed and predicted environmental changes in northern permafrost regions

As noted earlier, the greatest warming observed during recent decades and predicted for the next few decades is occurring at the highest latitudes in the northern hemisphere (IPCC, 2014; Kirtman et al., 2013; Serreze et al., 2000). Observed trends during the 20th century include reduced snow cover extent but increased annual precipitation; pronounced winter and spring warming; degradation of permafrost; and increased plant growth, greater abundance of shrubs, and northward migration of the tree line (IPCC, 2014; Min et al., 2008;

Fig. 4. The presence of permafrost along these cutbanks in the central Yukon River basin of Alaska increases bank resistance, as expressed in vertical banks between the overhanging mat of vegetation and roots in the active layer (top of bank) and the sloping bank toe where failure blocks accumulate. Banks in left and right views are approximately 3 and 4 m tall, respectively.

Peterson et al., 2002; Rawlins et al., 2010; Serreze et al., 2000). Although the magnitude of predicted changes varies among different warming scenarios, the common trends are: (1) increased precipitation and evaporation at high latitudes, with a net result of greater precipitation at 60 to 90°N; (2) warmer sea-surface temperatures in high latitudes and increased amounts of snow, but with a smaller fraction of precipitation falling as snow; (3) decreasing springtime snow cover extent; (4) greater mean surface warming over the Arctic during winter than global mean warming; (5) degradation of the upper 2–3 m of permafrost with a deepening of the active layer; and (6) increased runoff (Bintanja and Selten, 2014; Kirtman et al., 2013). In addition, models calibrated to contemporary conditions predict that Arctic rivers, which historically have had low suspended sediment loads but large interannual variations (Holmes et al., 2002), will likely have much greater suspended sediment fluxes to the ocean as climate warms (Gordeev, 2006; Syvitski, 2002). In this section, we describe changes that are occurring or will likely occur in the future in the factors that contribute to erosive force and erosional resistance. The factors that influence erosive force and erosional resistance operate over differing timescales, and timescales of change are partly dependent on the specific river segment under consideration. The climatic changes that are affecting and will affect these factors are also occurring over differing timescales. In the following discussion, we focus on the direction of change rather than rates of change.

4.1.1. *Permafrost*

Observed and modeled trends in permafrost characteristics suggest that erosional resistance will generally decrease along rivers with floodplain permafrost due to permafrost degradation. Model projections indicate that temperature will continue to rise at a faster rate in the high latitudes compared to the global mean temperature (IPCC, 2014). Permafrost temperatures have risen since the 1980s along with surface temperatures, and near-surface permafrost (within the top 3.5 m) is projected to decline by 37–81% by the end of the 21st century depending on future emissions (IPCC, 2014; Romanovsky et al., 2013; Slater and Lawrence, 2013). In some regions, widespread permafrost degradation is creating thermokarst features, particularly in ice-rich regions (Jorgenson et al., 2018; Jorgenson et al., 2006; Jorgenson et al., 2001).

Changes in active layer thickness are not spatially consistent across permafrost regions (Hinzman et al., 2013). In Eurasia, warming temperatures and increased snow depth, which provides insulation during the winter season, have resulted in a deepening of the active layer (Park et al., 2013; Zhang et al., 2005). In North America, decreased snow depths have exposed soils to colder temperatures in winter months, reducing active layer thickness in some locations (Park et al., 2013; Zhang et al., 2005), but there are other areas in North America where active layers have deepened (Osterkamp, 2005). Although changes in active-layer thickness are variable, it is clear that permafrost is

Fig. 5. River ice (left photo is from the bank of the Liard River) can gouge banks and topple and scar vegetation, as seen in the photo (right) of a bank of the Yukon River in interior Alaska. Photo on left taken by Chuck Blyth.

warming and degrading, and the trend will continue into the future (Hinzman et al., 2013; IPCC, 2014). This will lead to a decrease in erosional resistance.

4.1.2. Precipitation, runoff, and discharge

Observed and modeled trends in precipitation, runoff, and river discharge do not clearly indicate what the potential changes might be to erosive force exerted against banks and floodplain surfaces, but we hypothesize that the net effect may be an increase in erosive force and a decrease in erosional resistance. There has been an increase in fluxes of the freshwater cycle in the pan-Arctic region since the middle of the 20th century, with increases in river discharge, precipitation, and evapotranspiration, and these trends will likely continue into the future (Rawlins et al., 2010). Snow cover areal extent in spring has decreased since the 1970s in Arctic regions (Derksen et al., 2015), and it is likely that more precipitation will come as rain instead of snow in the future (Krasting et al., 2013; Nilsson et al., 2015). In addition, although highly variable, in general, snow thicknesses have increased (Hinzman et al., 2013). Discharge in permafrost zones is dominated by snowmelt, with peak flow corresponding to snowmelt in the spring. Particularly in boreal regions, secondary peak flows can occur in the summer due to convective rainstorms (Woo et al., 2008b). Hydrologic changes observed during recent decades include increased discharge from Canadian Arctic rivers (St. Jacques and Sauchyn, 2009) and increased baseflow in the Yukon River basin (Walvoord and Striegl, 2007). Although variations exist, total annual discharge has increased by an average of 9.8% since the 1970s across northern North America and Eurasia, with discharge in the first month of snowmelt increasing and the month in which peak discharge occurs decreasing slightly (Overeem and Syvitski, 2010). This increase has been attributed mainly to an increase in annual net precipitation in the pan-Arctic region forced by human-caused climate warming, which will likely continue into the future (Holland et al., 2006; Wu et al., 2005).

Changes in discharge patterns and magnitudes during the snowmelt period have also been detected. Snowmelt usually begins in April or May in many high-latitude rivers, with peak discharge in May or June (Overeem and Syvitski, 2010; Woo et al., 2008b). Trends show that peak discharge has decreased, snowmelt runoff has been arriving earlier in the year with earlier snowmelt, and the duration of high flows during snowmelt has increased (Goulding et al., 2009; Overeem and Syvitski, 2010; Semmens and Ramage, 2013; Woo et al., 2008b). Winter base flows have increased in most of the pan-Arctic, although a decreasing trend in winter-season flow in eastern North America is an exception (Rennermalm et al., 2010). Increased winter base flows have been linked to thawing permafrost, creating greater infiltration and longer subsurface flowpaths in the Yukon River Basin (Walvoord et al., 2012; Walvoord and Striegl, 2007) and in the Mackenzie River Basin (St. Jacques and Sauchyn, 2009).

Modeling efforts indicate that as warming increases into the next century, permafrost thaw will allow for the establishment of deeper flow pathways and increased groundwater discharge, although the response of groundwater discharge will likely be non-linear (Bense et al., 2009). In addition, there will be many variations in subsurface flow changes due to local subsurface characteristics (Walvoord and Kurylyk, 2016). Annual precipitation is likely to increase in the future (Koenigk et al., 2013), and there is the potential for an increase in the frequency of rainfall leading to flooding outside of the snowmelt period (Nilsson et al., 2015). In at least some river corridors, such as braided mountain streams (Mackay et al., 1973), summer storms can cause more significant floodplain changes than spring break-up or snowmelt floods.

In sum, trends indicate greater annual discharge, reduction in peak snowmelt discharge, a potential for longer duration of snowmelt flows, increasing base flow discharge, and the potential for greater and more severe flooding outside of snowmelt periods. The influence of these changes on erosive force is difficult to discern, as the reduction in peak snowmelt discharge reduces erosive force, but the longer duration of snowmelt, greater total annual discharge, increase in base flow discharge, and potential for flooding outside of snowmelt periods all increase erosive force. The influence of changes in discharge on erosional resistance is also somewhat unclear, but there may be a reduction in erosional resistance due to longer duration snowmelt and flashier flooding from rain as opposed to snowmelt.

4.1.3. Water temperature

Although not located in North America, insight into potential changes in water temperature can be gained through studies on the Lena River in Siberia. River water temperature has risen since the 1980s, which, along with reduced river ice thickness and an increase in spring breakup discharge, may explain accelerated erosion of the upstream end of vegetated islands during this period (Costard et al., 2003). In Svalbard, glacial-fed rivers exhibit lower temperature and lower temperature variations compared to snowmelt-fed and groundwater-fed systems, highlighting the potential for increased temperatures in Arctic rivers if glaciers recede and the relative contribution of glacial discharge decreases (Blaen et al., 2013). Although few studies exist on water temperatures, we expect that temperatures would increase with increasing surface temperatures. This would result in an increase in erosive force exerted on banks.

4.1.4. Icings and Ice jams

The extent and frequency of river icings are likely to decrease with warming climate (Pavelsky and Zarnetske, 2017). Morse and Wolfe (2015) found that winter warming intervals and autumn rainfall explained 28% of interannual variation in icing in Canada's Great Slave region during 1985–2014. They infer the potential for regionally different responses in future, with less frequent icings due to decreasing winter warming intervals in some regions, but increasing icings as a result of increasing autumn rainfall on the Canadian Shield, where the fill-and-spill runoff generation described by Spence and Woo (2003) creates strong threshold effects that favor icings formation. In Arctic Alaska, Pavelsky and Zarnetske (2017) found that the persistence and spatial extent of icings declined rapidly during 2000–2015. They attribute observed changes to either earlier and more intense melting or formation of less total icing volume.

Although somewhat uncertain, observations and predictions suggest a reduction in ice jam severity and frequency, likely reducing erosive force. Along with rising surface temperatures, the duration of ice cover on rivers has decreased, indicating later freeze-up and earlier break-up, and the thickness of ice cover has been reduced in some rivers (Beltaos and Prowse, 2009; Prowse et al., 2012). On the Peace River in Canada, modeling suggests that river ice extent and duration will continue to decline by the middle of the 21st century (Andrishak and Hicks, 2008). Higher air temperatures during spring will favor earlier breakup, but changes in breakup severity will likely vary regionally based on variations in winter precipitation and spring melt patterns: greater and more rapid snowmelt runoff could increase breakup severity, whereas smaller snowpacks with more protracted melt could decrease breakup severity (Prowse et al., 2006). Regions that experience a less dynamic breakup than previously due to thermal degradation of ice cover will likely have a smaller flood associated with break-up.

It is somewhat unclear what will happen in the future to the frequency and severity of ice jams due to the complicated nature of ice jam formation and the simultaneous future changes in snow cover and thickness and ice duration, thickness, and extent (Prowse et al., 2012). However, greater warming predicted in the future in permafrost regions indicates that ice may be thermally melted in place prior to mechanical break-up, potentially reducing ice-jam flooding (Prowse et al., 2012). In the Peace-Athabasca Delta, ice-jams have become less frequent since the 1960s, and modeling suggests that increased mid-winter thaws may reduce peak spring flows by reducing snowpack in the latter half of the 21st century, which will in turn reduce the frequency of ice-jam floods (Beltaos et al., 2006). Although the topic is understudied and complex, ice jam frequency and severity may be reduced in the future, resulting in reduced erosive force overall.

4.1.5. Sediment load

Across the Arctic and Subarctic, sediment load of rivers is predicted to increase as the climate warms (Syvitski, 2002). In the Mackenzie River Basin, there has been an increase in the delivery of suspended solids to the Beaufort Sea in the last 10 years, attributed to increased discharge and erosion within the drainage network (Doxaran et al., 2015). Degradation of permafrost leading to active-layer detachments and thaw slumps can extend the drainage network, increase sediment loads in runoff and in streams, and enhance deposition downstream (Lamoureux and Lafrenière, 2009; Toniolo et al., 2009). On the North Slope of Alaska, thermokarst hillslope failures have increased since the 1980s, likely increasing sediment loads to streams (Gooseff et al., 2009). Thaw slumps in the Mackenzie River watershed in the Northwest Territories of Canada can greatly increase sediment loads within streams and rivers at a variety of watershed scales compared to streams unaffected by thaw slumps (Kokelj et al., 2013). With rising temperatures and the potential for longer duration high flows causing greater fluvio-thermal erosion (Costard et al., 2014), degradation of bank permafrost may increase, adding to the sediment load of rivers. Increases in bed-material load will likely increase erosive force as deposition of bars causes flow deflection towards bank. However, large-scale increases in sediment supply may overwhelm the discharge capacity to transport the sediment, reducing erosive force relative to sediment supply. Thus, changes in sediment supply could either increase erosive force or reduce erosive fore relative to incoming sediment.

4.1.6. Vegetation

Vegetation has also been changing in permafrost regions in a manner that suggests a net effect of increasing erosional resistance. Changes in temperature, precipitation, soil moisture, and thawing depth will likely lead to changes in vegetation cover, with woody vegetation likely to be favored as the active layer deepens, particularly on well-drained sites such as river banks (Lloyd et al., 2003). General trends in vegetation include an increase in above-ground biomass in the Arctic over the last 30 years (Epstein et al., 2012), shrub expansion in both valley bottoms and hillslopes (Tape et al., 2006), and treeline advance since 1900 at a majority of study locations across the Arctic region (Harsch et al., 2009). Treeline will likely continue advancing into the latter half of the 21st century, and there will likely be an expansion of tall shrubs in the tundra (Zhang et al., 2013). As temperature has warmed, there has been an increase in woody vegetation along the transition between the boreal forest biome and the tundra biome in some places (Urban et al., 2014). Because active-layer depth limits the growth and rooting depth of plants (Grant, 2015), increasing active-layer depths could lead to increased root growth and rooting depth. Advances in treelines, increases in active-layer depths, and an increase in the extent of woody vegetation will likely increase erosional resistance.

Changes in vegetation cover may also be driven by changes in wildfire. Using historic fire records for the North American boreal region during 1959–1999, Kasischke and Turetsky (2006) documented a doubling of annual burned area and more than a doubling of the frequency of larger fire years, while the proportion of total burned areas from human-ignited fires decreased.

4.1.7. Freeze/thaw cycles

Trends in the frequency of freeze/thaw cycles are somewhat uncertain, but it is likely that temperature variations and warming will result in more frequent crossings of 0 °C within soils, particularly for lower latitudes within permafrost regions (Henry, 2008; Nilsson et al., 2015). This would result in more freeze/thaw cycles throughout the year (Henry, 2008), which will reduce erosional resistance.

4.1.8. Large wood loads

Past trends in large wood loads have not been studied in permafrost regions, nor have future trends in wood loads been predicted. Because wood is recruited to channels through mass movements and bank erosion (Martin and Benda, 2001; Ott et al., 2001), we infer that wood loads will increase with increases in hillslope instability and bank instability caused by thermokarst development and permafrost degradation. Increased dispersed LW loads will likely decrease erosive force and increased LW jam load and number will have varying effects on erosive force. The net effect of these combined changes is uncertain.

4.1.9. Grain size

Because grain size is dependent on regional variations in geology, geomorphic setting, valley geometry, and geologic history, changes in grain size with climate change cannot be generalized. Site-specific changes will reflect the combined effects of changes in factors such as the relative importance of discrete source areas (e.g., hillslopes affected by thermokarst-induced failure, or channel banks affected by permafrost degradation) and sediment transport and marginal and overbank deposition.

4.1.10. Complexities and summary of potential changes

Table 3 summarizes the factors affecting erosive force and erosional resistance and the predicted general direction of change in these factors associated with climate warming. Uncertainties exist, and we acknowledge that multiple factors changing at once will likely have different resulting effects on erosive force and erosional resistance. For example, changes in vegetation will impact hydrologic and geomorphic processes. More extensive vegetation cover is likely to reduce upland erosion rates during snowmelt and rainfall (Forbes and Lamoureux, 2005). Vegetation can modify expected channel planforms. Huisink et al. (2002) attribute the absence of braided channel planform in the Usa River catchment of Russia to the presence of dense vegetation cover that limits sediment supply and prevents braiding, despite the presence of discontinuous permafrost and locally steep slopes. Vandenbergh (2001), in classifying channel planform types in periglacial regions, also notes the absence of braided channels in regions with continuous vegetation cover. Where steep uplands supply sediment to continuously vegetated lowlands, the lowland channels are likely to be stable, high-energy, anabranching systems (Vandenberghe, 2001). The specific trajectories of vegetation change matter, however. Permafrost degradation in the Mackenzie River drainage, for example, is changing the distribution and proportion of the three main cover types of black spruce (*Picea mariana*) forest underlain by permafrost, channel fens with a floating peat mat, and flat bogs, each of which has distinct hydrological properties (Quinton et al., 2011). Permafrost thaw can also transform forests to fens or bogs. Average annual basin runoff correlates positively with the percentage of the basin covered by fens and negatively with the percentage covered by bogs (Quinton et al., 2011; Quinton et al., 2009), indicating that the interacting effects of vegetation and runoff changes depend on trajectories of change. In terrain with higher relief, characteristics such as elevation and aspect create local spatial heterogeneity in vegetation change (Buma and Barrett, 2015).

The effects of these predicted changes will likely vary between specific sites depending on the magnitude of change in processes and feedbacks that arrest or enhance changes. For example, thinner snow packs could reduce insulation of permafrost during winter and limit increases in thaw depth (Walvoord and Kurylyk, 2016). Change from continuous to discontinuous or sporadic permafrost and increases in the formation of taliks could increase soil moisture storage capacity and change runoff and groundwater processes in complex ways (Walvoord and Kurylyk, 2016). Changes in storm intensity and precipitation could shift peak flows from spring snowmelt to summer rainfall. Dugan et al. (2009) report a case study from a small Arctic watershed in which exceptional summer temperatures during July 2007 caused early, deep

Table 3
Summary of the factors affecting erosive force and erosional resistance and the predicted general direction of change associated with climate warming.

Erosive force	Change in erosive force with increase in factor	Direction of change in factor due to warming	Likely effect of warming on erosive force due to change in factor
Discharge (Duration, Magnitude)	+	+	+
Water temperature	+	+	+
Sediment load	±	+	±
Dispersed Large Wood			
Stationary	-	+	-
In transport	+	+	+
Large Wood Jams			
Upstream	-	+	-
At	±	+	±
Downstream	±	+	±
Ice Jams			
Upstream	-	-	+
At	±	-	-
Downstream	+	-	-
Erosional resistance	**Change in erosional resistance with increase in factor**	**Direction of change in factor due to warming**	**Likely e ffect of warming on erosional resistance due to change in factor**
Vegetation	+	+	Unknown
Buried logjams	+	Unknown	Unknown
Permafrost	+	-	-
Discharge (Duration, Magnitude, Variation)	-	+	-
Freeze/Thaw	-	+	-
Grain size	variable	Variable	Variable

permafrost thaw, which in turn resulted in substantially elevated rainfall runoff and suspended sediment load during a rainfall event of average magnitude and intensity. Obviously, the detailed effects of climate warming will vary between specific sites and through time. Permafrost degradation or persistence will likely depend on site-specifics characteristics. For example, permafrost underlying rocky uplands degrades after wildfire and soils become well drained over a period of decades, but permafrost persists after fire in silty uplands (Jorgenson et al., 2013).

In addition, we recognize that thresholds of change exist within geomorphic systems, and that responses to climatic changes may be non-linear and complex (Church, 2002; Schumm, 1979). Knowledge of contemporary hydrologic processes in permafrost regions (Forbes and Lamoureux, 2005); Loisel et al., 2017; Myers-Smith et al., 2008; Prowse et al., 2006; Vandenberghe, 2001) suggests that thresholds are likely to characterize upland and river responses to changing temperature and precipitation patterns. For example, Avis et al. (2011) predict that, following an initial increase in wetland area within permafrost zones as a result of the lengthening of the thaw season and greater surface moisture, the number of wetland-conducive days and the extent of wetlands will decline dramatically as the active layer deepens and subsurface drainage develops.

The magnitude and rates of change, as well as the details of interactions among diverse factors, will undoubtedly vary between specific river segments and river networks. However, providing general predictions for how changing factors will influence erosive force and erosional resistance allows us to hypothesize a net effect of climate change on channel and floodplain form and process.

5. Conceptual models of predicted changes in floodplain process and form

Although it is difficult to determine the cumulative effects of the changes in factors described in Section 4 on erosive force and erosional resistance, we present hypotheses regarding the cumulative effects in the form of conceptual models. We provide these conceptual models as frameworks that may apply to different river corridors in North American permafrost regions. We present two contrasting scenarios, in which the ratio of erosive force to erosional resistance either increases or decreases, and then discuss the net effects of these two scenarios on channel process and form and floodplain process and form. We then discuss regional differences that may influence the effects of changing erosive force and erosional resistance.

Although we discuss two scenarios below in which the ratio of erosive force and erosional resistance changes directionally, we acknowledge that some river corridors may not experience any change in the ratio of erosive force and erosional resistance as a result of climate warming. This could result from lack of modifications of erosive force and erosional resistance, or relatively equal changes in erosive force and erosional resistance (both increase, or both decrease). Although these options are possible, the greatest modifications to process and form in river corridors will likely result from changes in the ratio of erosive force and erosional resistance. Thus, we focus on the two scenarios in which the ratio changes.

5.1. Increase of erosive force relative to erosional resistance

The first scenario that we discuss is the potential for an increase in the ratio of erosive force to erosional resistance. This scenario proposes

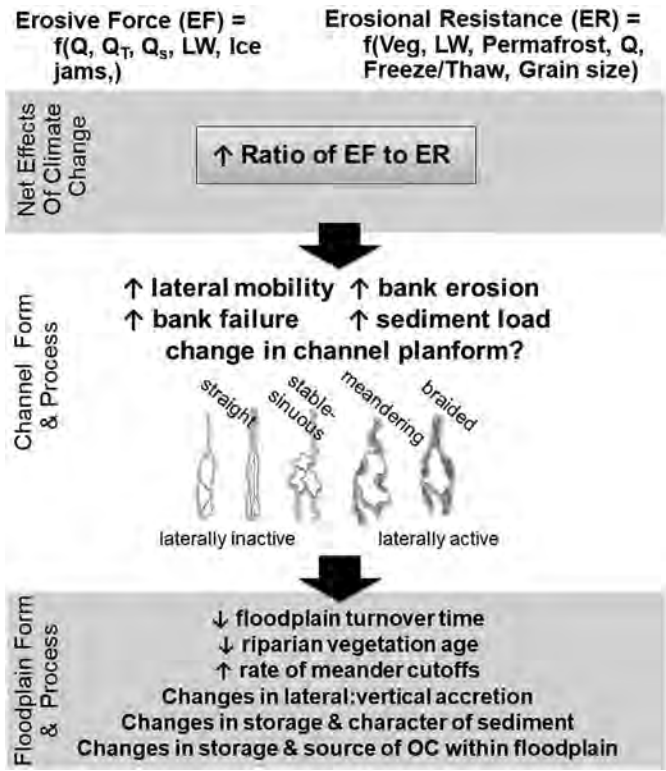

Fig. 6. Conceptual model predicting one scenario for climate change effects on channel form and process and floodplain form and process for rivers in permafrost regions. The ratio of erosive force to bank resistance could increase due to increasing discharge and a reduction in permafrost extent, resulting in increased lateral migration rate, bank erosion, and bank failure. The changes would also result in changes in floodplain form and process. Schematic of planform types within channel process and form section is after Nanson and Knighton (1996).

that, although the cumulative impact on erosive force is somewhat uncertain due to the many factors influencing erosive force, the net effect could be increased erosive force as a result of increased discharge. The reduction of the ratio of erosive force to erosional resistance could also result from the loss of near-surface permafrost and reduction in areal permafrost extent. Because permafrost strongly increases bank cohesion, permafrost thaw will likely have the greatest impact on erosional resistance relative to the other factors. The increase in the ratio of erosive force to erosional resistance would result in cascading effects for channel process and form and thus floodplain process and form (Fig. 6).

The reduction in erosional resistance relative to erosive force would likely change channel process and form by some combination of bank erosion, bank failure, lateral migration, channel width, and sediment recruitment. This would affect channel type, with all channel types becoming wider and/or more laterally active. The potential also exists for some channel types to shift towards more laterally active planform types. One end member of a spectrum of river types in terms of lateral activity is a straight or, more commonly, a stable-sinuous river (Nanson and Knighton, 1996). At the other end of the spectrum is a braided channel, which can be highly laterally active. This spectrum exists for both single-channel rivers and anabranching rivers (Nanson and Knighton, 1996). Thus, in this scenario, straight or stable-sinuous rivers would meander more, potentially becoming fully meandering. Meandering rivers could become more braided, and braided rivers may have higher rates of avulsions or the development of more channels within the braid plane.

Warming and resulting changes would also be reflected in floodplain form and process. It is important to note that floodplains along many river segments reflect past conditions of formation (Dunne and Aalto, 2013), and thus the time scale for adjustment in floodplain form to climate change may be longer than the time scale for channel adjustment. However, increases in lateral migration rate and bank erosion would likely result in a decreased floodplain turnover time, which is the average time it takes for a channel to occupy the entirety of its geologic floodplain. With a decrease in floodplain turnover time, the average age of riparian vegetation could decrease. If channel erosion and deposition are more quickly re-working previously stable surfaces, riparian vegetation will be disturbed more frequently, although age of riparian vegetation also reflects other disturbances such as wildfire. There is also the potential for an increased rate of meander and chute cutoffs and an increased rate of oxbow lake formation with increased lateral channel mobility.

The relative amounts of lateral and vertical accretion in floodplains would also be affected, with resulting changes in the character and storage of floodplain sediment. Increased lateral migration could result in more lateral accretion, which would be compounded by increased sediment supply due to bank collapse and upland sediment delivery and even more lateral accretion. Because lateral accretion deposits are coarser than vertical accretion deposits (Wolman and Leopold, 1957), greater lateral accretion could result in coarsening of portions of the floodplain. However, if bank resistance is reduced over the long term due to permafrost melt, there could potentially be a reduction in bank heights, allowing for the ability of lower flows to access floodplain farther from the river and promoting increased vertical accretion. An increased rate of oxbow lake formation provides backwater areas available for filling via vertical accretion, potentially resulting in a greater proportion of the floodplain being composed of finer sediment.

The storage capacity of organic carbon (OC) within floodplains would also change, as well as the source of OC within floodplains. If the source of floodplain OC is primarily downed wood from the floodplain forest, leaf litter, and organic layers developed on stable floodplain surfaces (autochthonous organic carbon), a faster lateral migration rate could result in lower floodplain OC storage. Limited studies of floodplain OC storage suggest that autochthonous material is the dominant source of OC on floodplains (Lininger et al., 2018; Sutfin et al., 2016).

If the source of floodplain OC is primarily from river deposition of particulate OC along with sediments or OC adsorbed onto fine sediments (allochthonous organic carbon), changes in lateral and vertical accretion will also influence the amount of floodplain OC. The greatest effect is likely to result from increased lateral migration rates that affect OC stored as wood within the channel and floodplain. Greater bank erosion would result in greater recruitment of wood, and the increase in lateral mobility could increase the amount of wood buried in sediments within the floodplain as the actively meandering or avulsing channel leaves dispersed or concentrated wood on bars that become attached to the floodplain (Collins et al., 2012; O'Connor et al., 2003).

Changes in floodplain organic carbon storage are also likely as a direct result of permafrost thaw. For example, floodplains could become less saturated as the impermeable permafrost layer degrades, sub-surface flow paths increase, and infiltration increases. Decreased saturation could lead to more oxidation of organic carbon because saturated conditions inhibit microbial decomposition and processing of organic matter (Trumbore and Czimczik, 2008). Decomposition of previously frozen organic matter that was unavailable to microbes within permafrost may reduce floodplain carbon storage overall (Drake et al., 2015; Schuur et al., 2008). The combined effect of release of OC from melting permafrost and drier floodplains and declining storage of autochthonous floodplain OC could decrease floodplain OC storage. However, many uncertainties exist. Plant macrofossil records of change during the past century at a floodplain site in Alaska indicate the difficulty of exactly predicting the direction of future change: future warming and/or increased wildfire disturbance at this site could promote permafrost degradation, expansion of peatland at the expense of black spruce forest, and increased carbon storage, or increased drought

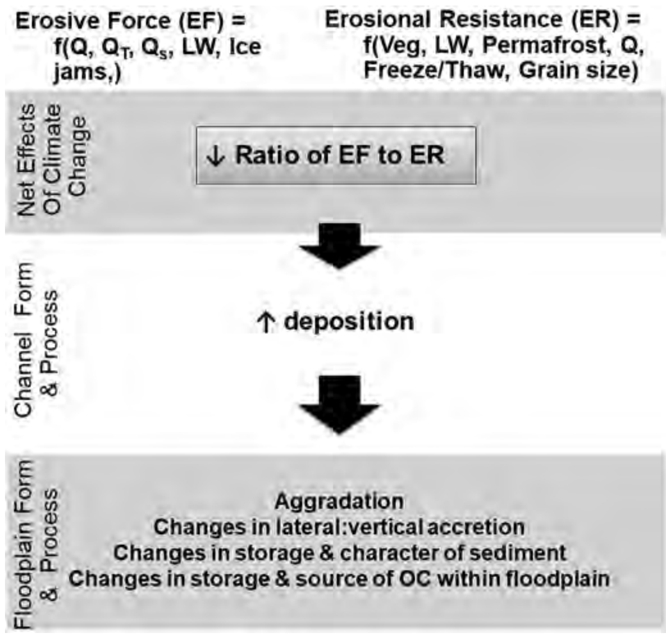

Fig. 7. Conceptual model predicting one scenario for climate change effects on channel form and process and floodplain form and process for rivers in permafrost regions. The ratio of erosive force to bank resistance could decrease due to reduction in erosive force relative to increased sediment supply from enhanced thermokarst activity and bank failure. The changes would also result in changes in floodplain form and process.

could reduce the growth of both black spruce and peatland, leading to decreased carbon storage (Myers-Smith et al., 2008).

5.2. Decrease in erosive force relative to erosional resistance

Another possible outcome of warming on river and floodplain dynamics is a decrease in the ratio of erosive force to erosional resistance, with cascading effects for channel and floodplain process and form (Fig. 7). This change could occur in part due to the reduction in erosive force relative to sediment supply. Sediment supply could increase due to increased bank failure from loss of near-surface permafrost and reduction in permafrost extent and increased inputs from thermokarst activity such as slumps (Section 4.1.5). Some of the effects of the decrease in the ratio of erosive force to erosional resistance are the same as described in the above scenario (Section 5.1), so this section focuses on the effects that differ from the above scenario. However, common changes that will likely occur in both scenarios are briefly described in Section 5.3.

The reduction in the ratio of erosive force to erosional resistance would affect channel form and process by primarily increasing deposition within the channel. This would result in aggradation within the channel and floodplain, although the type and location of sediment deposited would likely vary depending on specific locations. For example, if sediment characteristics are primarily fine-grained, there could be increased vertical accretion of fines, but if sediment within a river corridor is coarser, there could be increased lateral accretion. In general, aggradation, changes in the ratio of lateral to vertical accretion, and changes in the storage and character of sediment would result from enhanced sediment supply and a reduction in the ratio of erosive force to erosional resistance.

Similar to the first scenario presented, a reduction in the ratio of erosive force to erosional resistance would likely cause changes in the storage and source of OC within the floodplain. Increased deposition and burial of OC could enhance floodplain OC storage within both soil and large wood, since burial can protect OC from decomposition (Doetterl et al., 2016). Increased deposition and delivery of fines to floodplains could also enhance floodplain soil OC storage, since fines are associated with higher OC concentrations (Pinay et al., 1992). However, because soil OC in floodplains is controlled by multiple geomorphic influences, such as grain size, soil moisture, and geomorphic unit (e.g., bar, backwater wetland, swales, etc.) (Lininger et al., 2018), an increase in floodplain aggradation may not result in an overall increase in floodplain OC storage.

5.3. Common likely changes in both scenarios

Although the origins and mechanisms of the two scenarios presented above are different, there are similarities in the resulting changes in channel and floodplain process and form. Both scenarios would enhance sediment supply and thus deposition in the river corridor, with resulting changes to floodplain OC storage. These inferences are supported by changes that occurred to rivers and floodplains at diverse sites during the Pleistocene-Holocene transition in regions now well south of permafrost zones and by past river responses in permafrost regions during periods of warming (Mann et al., 2010; Murton and Belshaw, 2011), in which aggradation occurred.

5.4. Regional differences

The effects of climate change on river corridor dynamics depend on regional characteristics (Ashmore and Church, 2001) not explicitly addressed in the general scenarios discussed in Sections 5.1 and 5.2. Exceptionally high sediment yields can follow major rainfall in some Arctic catchments (Cogley and McCann, 1976), for example, whereas suspended sediment transport only occurs during snowmelt in other catchments (Forbes and Lamoureux, 2005). In this section, we discuss regional differences in valley geometry, catchment relief, bedrock topography, thermokarst activity, climate, vegetation cover, and disturbance regime (e.g., wildfires).

5.4.1. Valley geometry

Valley geometry here refers to the extent to which a floodplain is developed, as well as the valley gradient and the lateral coupling between uplands and the active channel. These characteristics tend to be correlated in that channels with extensive floodplains are likely to be of lower gradient and less coupled to the adjacent uplands because of the buffering created by the presence of a floodplain (Church, 2002). Drainage area can influence valley geometry, with smaller basins tending to have smaller floodplain extents and closer coupling with hillslopes. An example of differences in river dynamics between channels with differing valley geometry comes from watersheds in the Kuparuk River basin on the North Slope of Alaska (Kane et al., 2008). The smaller, higher gradient channels are most affected by summer rainfall-generated floods, which can have peak flows three to four times greater than snowmelt peak flows. In contrast, larger, lower gradient channels on the coastal plain are most affected by snowmelt peak flows, not least because the largest summer rainfall floods are only about a quarter of the magnitude of the snowmelt peak flows (Kane et al., 2008). Although continuous permafrost and associated limitations on near-surface storage in the shallow active layer contribute to the response of the small catchments to rainfall, this pattern of smaller catchments being more geomorphically responsive to localized summer rainfall than larger catchments is also well documented for drainage basins in temperate latitudes (O'Connor et al., 2002). In addition, changes in ice jam flood timing have been more extreme in small river basins compared to large river basins, indicating that small river basins may be more sensitive to changes in ice jam floods (Rokaya et al., 2018). Glacial history can also influence valley geometry, with wider U-shaped valleys with potentially larger floodplains in glaciated mountain valleys when compared to unglaciated valleys (Montgomery, 2002).

Valley geometry can also influence permafrost distribution. As discussed in Section 2.2, the presence of permafrost can increase hydrological responsiveness to precipitation inputs, although the details of this responsiveness vary substantially in relation to small-scale (sub-meter) surface topography, vegetation cover, and permafrost extent. Floodplains on Arctic and Subarctic rivers can be underlain by permafrost, although floodplain wetlands and secondary or recently abandoned channels are likely to contain thicker active layers or be underlain by unfrozen zones. For example, taliks (unfrozen zones within permafrost) many times forms beneath surface water bodies (Walvoord and Kurylyk, 2016), thus the presence of negative relief and floodplain water bodies in extensive floodplains can influence the distribution of permafrost and subsurface flows. Unfrozen zones can also occur below rivers, such as the Yukon River (Minsley et al., 2012). The details of floodplain stratigraphy, permafrost spatial extent and active layer depth, and surface topography are likely to create differences in contemporary river process and form among diverse river segments, as well as differences in response to warming climate.

The presence of an extensive floodplain can influence the relative importance of upland versus river corridor sources of sediment and organic carbon and sediment concentration and yields. The Yukon has extensive floodplain wetlands, whereas those along the Mackenzie are restricted to limited segments of the river's course. As floodplain extent increases, river corridor sources of sediment, as opposed to upland sources, can become progressively more important (Meade, 2007), even if the floodplain acts as a long-term sediment sink (Mertes et al., 1996). Floodplains and the lateral coupling of river corridors and uplands also influence hydrologic flow paths to the active channel, which can be surface or subsurface pathways in both uplands and the river corridor. Large river floodplains can be geomorphically complex with variable topography, which can influence the preferential movement of water through the subsurface and surface (Dunne and Aalto, 2013; Lewin and Ashworth, 2014).

Extensive alluvial floodplains on large rivers with high banks may respond differently to near-surface permafrost degradation compared to smaller rivers with less overall sediment storage and shorter banks. On large rivers with high banks, permafrost degradation in the top 1–3 m may not influence erosional resistance as much as along rivers in which the entire bank height is degraded and thawed.

The presence of a wide valley and extensive floodplain can influence channel planform by facilitating greater sinuosity and a wider belt of anastomosing or braiding in multithread channels. The characteristics of the channel planform will govern the length of channel bank per unit length of valley, as well as the manner in which floods and floating ice shape the channel boundaries. Channels with islands and bars or other irregularities that preferentially accumulate ice, for example, can be protected by accumulation of ice during the initial stage of breakup but subsequently experience enhanced erosive forces during seasonal peak flows (Costard et al., 2014).

5.4.2. Catchment relief, bedrock topography, and potential for thermokarst

Relief can influence characteristics such as type and size of slope instability associated with permafrost degradation (Fischer et al., 2006). Bedrock topography refers the characteristics of underlying bedrock within the river corridor and connected uplands. Bedrock topography is important as it controls the depth of the active layer and permafrost. Where bedrock is close to the surface, the effects of warming climate may be less pronounced because the bedrock may be able to continue to create a mostly impermeable underlying layer (Devito et al., 1996). Depth to bedrock varies regionally across the Canadian Shield, creating spatial variations in permafrost and associated groundwater storage and runoff generation (Morse and Wolfe, 2015; Spence and Woo, 2003) that influence contemporary river process and form and will influence responses to warming climate.

One of the more important regional differences is the presence of thermokarst and retrogressive thaw slumps in areas with thick sediment deposits and substantial ground ice, which causes thermokarst features when thawed (Olefeldt et al., 2016). In addition, rivers in higher relief terrain containing thermokarst are more connected to the sediment source provided by thermokarst. Sites with slumps exhibit diurnal fluctuations in water, solute, and sediment flux, as well as suspended sediment and solute concentrations several orders of magnitude higher than rivers in regions not affected by slumps (Kokelj et al., 2013).

5.4.3. Climate

Climate is mentioned here in acknowledgement of the differences in mean annual temperature and associated permafrost characteristics, as well as differences in precipitation amount, type, and temporal distribution within the North American permafrost region. Mean annual temperature is zoned broadly by latitude in the North American permafrost region, with coldest temperatures farthest north. Sporadic and discontinuous permafrost along the southern boundaries of the region is likely to disappear most rapidly, so that changes in factors that affect river dynamics will be time transgressive from south to north. At more local scales, elevation- and aspect-related spatial heterogeneity in temperature will undoubtedly create differences in the rate of permafrost degradation and accompanying changes.

5.4.4. Vegetation cover

As with climate, contemporary vegetation cover varies substantially across the North American permafrost region, both within a relatively small watershed of a few tens of kilometers, and across the region as a whole. Basic categories of vegetation include dry, moist, and wet tundra (grasses, sedges, lichens, woody shrubs) and boreal forest (*Picea* spp., *Pinus* spp., *Populus* spp., *Betula* spp.) (Bailey, 1995). Spatial characteristics of vegetation can influence surface runoff (Young et al., 1997), susceptibility to wildfire (see Section 5.3.5), and the erosional resistance of the floodplain surface and channel banks (section 4.1.3). Consequently, as with many of the other factors discussed in this paper, the details of vegetation cover influence contemporary river process and form and ongoing responses to warming climate.

5.4.5. Wildfire regime

Wildfires can alter soil moisture and temperature, organic matter accumulation, and vegetation communities and thus thermal characteristics of the ground surface and shallow subsurface. Natural fire recurrence intervals in boreal forest range from 50 to 200 years, to a high of 500 years in moist regions of eastern Canada (Bonan and Shugart, 1989). Regions with low precipitation and high summer temperatures, such as the Yukon Flats region of the Yukon River drainage, have frequent fire disturbance, with recurrence intervals of 37 to 166 years (Drury and Grissom, 2008).

Most tree species of the boreal forest except for pines are not fire resistant, so fires result in the death and regeneration of forest stands, but no single sequence of revegetation follows fire (Bonan and Shugart, 1989). In boreal forests, fire can result in increases in the thermal conductivity of the ground and decrease surface albedo, allowing for greater absorption of insolation, but these effects depending on the extent of burning of the surficial organic layer (Yoshikawa et al., 2002). If the organic layer thickness is reduced by the fire, thickness of the active layer increases in subsequent years and this change may persist for more than a decade (O'Donnell et al., 2011; Yoshikawa et al., 2002). Changes in fire regimes will likely influence vegetation, permafrost, and large wood loads along North American permafrost rivers.

6. Broader implications

The changes in floodplain form and process hypothesized above have implications for sediment yields, nutrient and OC export to the Arctic, aquatic and riparian habitat, and infrastructure in permafrost regions. Because floodplains are sites of deposition, erosion, and processing of nutrients, changes in floodplains will directly impact the exports of sediment, OC, inorganic nutrients, and dissolved ions to the Arctic Ocean. Sediment flux is predicted to increase in high latitude rivers as climate warms (Syvitski, 2002). The timing, form, and amounts of nutrients and OC exported to the Arctic Ocean will also be altered (Frey and McClelland, 2009; Guo et al., 2007; Holmes et al., 2013), although trajectories of change may differ among rivers (Holmes et al., 2012).

In Alaska and the Yukon Territory, watersheds with permafrost export more DOC compared to similar watersheds that are permafrost-free (Frey and McClelland, 2009; Petrone et al., 2006; Walvoord and Striegl, 2007). This is attributed to leaching of organic matter from the organic-rich surface layer in the surface-runoff dominated permafrost zones. A reduction in DOC export to the Arctic Ocean is predicted in these regions as infiltration increases and more DOC adsorbs onto mineral grains (Frey and McClelland, 2009; Walvoord and Striegl, 2007). There also may be an increase in POC export as bank erosion increases along with permafrost thaw (Guo et al., 2007).

Degradation of permafrost via thermokarst development can also directly influence DOC fluxes. Thermokarst features can preferentially form on channel banks and lakeshores on Alaska's North Slope region (Abbott et al., 2014). Abbott et al. (2015) distinguish slides, thaw slumps, and gullies resulting from thermokarst. DOC concentrations in river flow increase substantially where thermokarst is present, but the persistence of these increases varies. Slides form suddenly and stabilize within the same season. Large thaw slumps commonly remain active for a few decades and gullies can remain active for a decade, so slumps and gullies have the most persistent effects on DOC fluxes (Abbott et al., 2015).

The Arctic Ocean comprises only 1% of global ocean volume, but receives approximately 11% of global freshwater discharge (McClelland et al., 2012). Terrestrial exports of organic matter provide an integral input to food webs in near-shore Arctic environments, supporting benthic organisms and thus many fish and bird species (Dunton et al., 2012). Changes in the exports of nutrients and organic matter, and the timing of those exports, will likely have broad effects on productivity within the Arctic Ocean (McClelland et al., 2012).

Floodplains will mediate changes in sediments and nutrients as these materials move through the drainage network. Globally, estimates for the amount of carbon buried in sediments in freshwaters ranges from 0.2 to 1.6 Pg C yr^{-1}; this large range in estimates demonstrates the lack of information available on the burial of carbon within floodplains, lakes, and reservoirs (Regnier et al., 2013), highlighting the need to incorporate floodplain dynamics into estimates of nutrient and organic matter fluxes.

Changes in the export of sediment to the Arctic Ocean are important in their own right and because of the nutrients and contaminants that can travel adsorbed to silt and clay. Fine sediment and adsorbed materials are incorporated into Arctic Ocean sea ice when fine sediment remains suspended in the water column, a common occurrence in the shallow shelf zone where wave dynamics keep the sea bottom agitated (Pfirman et al., 1995). The sediment can travel with the sea ice throughout the Arctic basin and into the North Atlantic Ocean through the Fram Strait. River transport contributes a major proportion of the sedimentary and geochemical budgets for the Arctic Ocean (Standring et al., 2008).

The changes in floodplain process and form with warming will alter physical structure and nutrient cycling in floodplain environments, impacting aquatic and riparian habitat provided by floodplains. Secondary channels within floodplains and floodplain lakes can provide important rearing areas for juvenile fish species (Bellmore et al., 2012). Canada's boreal forest, much of which lies within the Mackenzie River drainage, supports an estimated three to five billion migratory birds, from warblers to whooping cranes, and a higher diversity of breeding birds than anywhere else in North America (Dyke, 2000). A majority of the world's wetlands occur between 50 and 70°N, and many of these are on floodplains. Floodplain lakes with differing amounts of river inflow have different water chemistry and aquatic communities. In the Mackenzie drainage, turbid water in frequently flooded lakes limits the growth of rooted macrophytes and algae, whereas lakes with less river connectivity have clear water and total biomass comparable to temperate and tropical floodplain lakes, despite cold temperatures and a short growing season (Brunskill et al., 1973). Similarly, naturally turbid rivers support less abundant and diverse invertebrate fauna than clearwater rivers in the Mackenzie (Brunskill et al., 1973). The accelerated hydrologic cycle and floodplain turnover that could potentially occur as climate continues to warm will thus alter not only river corridor morphology, but also nutrient retention, habitat, and biodiversity and abundance for aquatic and riparian environments in ways not yet understood.

Changes in channel form and process and associated changes in floodplain form and process will also influence infrastructure and the ability of people to travel in the Arctic and Subarctic. Floodplains underlain by permafrost in North America are primarily characterized by widely dispersed infrastructure in the form of roads and bridges and relatively small settlements. However, many settlements are along rivers and located within floodplains, and permafrost thaw and degradation can create extensive damage to infrastructure built along river banks (Nelson et al., 2002). In addition, infrastructure that cross channels and floodplains, such as pipelines, could be affected by permafrost degradation and changing channel and floodplain processes. A reduction in ice cover on rivers and in permafrost extent would reduce the availability and use of winter roadways, because transportation and travel commonly occurs on frozen water bodies, frozen rivers, and frozen ground during the winter months (Stephenson et al., 2011).

7. Conclusion

We have provided hypothesized conceptual models for changes in floodplain form and process in permafrost regions as climate change continues to alter permafrost extent, precipitation patterns, runoff, and other factors that influence erosive force and erosional resistance. The specific hypotheses and conceptual models presented here will be testable with time within diverse river corridors. However, there are reasons to start thinking now about potential impacts of warming on floodplains in permafrost regions, ranging from global-scale influences on OC dynamics to regional effects on terrestrial and riparian wildlife, nearshore biota, and human communities and infrastructure.

We present two scenarios, one in which the ratio of erosive force to erosional resistance will increase, and one in which the ratio will decrease. If the ratio of erosive force to erosional resistance increases, resulting changes could include more bank erosion and failure, lateral migration, and sediment supply. This could cause a shift in river types, with stable channels, such as stable-sinuous, moving towards more active channel types, such as meandering or braided channels. These changes would have cascading effects on floodplain form and process, including causing decreased floodplain turnover time, decreased riparian vegetation age, changes in the relative importance of lateral and vertical accretion, increased rate of meander cutoffs, changes in the nature and storage of sediment, and changes in the source and storage of OC within floodplains. If the ratio of erosive force to erosional resistance decreases, resulting changes could include increased deposition within river corridors, causing aggradation, changes in the proportion of vertical to lateral accretion, and changes in the source and storage of OC within floodplains.

Many rivers in North American permafrost regions are still relatively un-altered by humans. However, with the continued emissions of fossil fuels and increased warming that will ensue, we will significantly alter floodplain form and process in permafrost regions. There are many studies describing changes in exports to the Arctic Ocean and hydrologic changes as a result of climate warming, but very little attention has been paid to changes in floodplains and the resulting effects of those changes. Rivers are far from the ‘neutral pipes’ conceptualized in early studies of global carbon dynamics (Cole et al., 2007). Because floodplains are sites of sediment and organic matter storage and nutrient processing, they affect the movement and behavior of water and dissolved and particulate nutrients in ways that influence both the river corridor and downstream areas, including delta and marine environments. Warming climate is already causing documented changes in hydrological processes in regions underlain by permafrost, but conceptual and numerical models of these changes have not yet effectively incorporated channel-floodplain connectivity and lateral fluxes, despite the significant influence of such connectivity in rivers with extensive floodplains. This underscores the need to document and interpret floodplain dynamics in permafrost regions as we continue to alter global climate.

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References

Aalto, R., Lauer, J.W., Dietrich, W.E., 2008. Spatial and temporal dynamics of sediment accumulation and exchange along Strickland River floodplains (Papua New Guinea) over decadal-to-centennial timescales. J. Geophys. Res. 113, 1–22. https://doi.org/10.1029/2006JF000627.

Abbott, B.W., Larouche, J.R., Jonew, J.B., Bowden, W.B., Balser, A.W., 2014. Elevated dissolved organic carbon biodegradability from thawing and collapsing permafrost. J. Geophys. Res. Biogeosci. 119, 2049–2063. https://doi.org/10.1002/2014JG002678.

Abbott, B.W., Jones, J.B., Godsey, S.E., Larouche, J.R., Bowden, W.B., 2015. Patterns and persistence of hydrologic carbon and nutrient export from collapsing upland permafrost. Biogeosciences 12, 3725–3740. https://doi.org/10.5194/bg-12-3725-2015.

ACIA, 2005. Arctic: Arctic climate impact assessment. Cambridge University Press, New York.

Aiken, G.R., Spencer, R.G.M., Striegl, R.G., Schuster, P.F., Raymond, P.A., 2014. Influences of glacier melt and permafrost thaw on the age of dissolved organic carbon in the Yukon River basin. Glob. Biogeochem. Cycles 28, 525–537. https://doi.org/10.1002/2013GB004764.

Allmendinger, N.E., Pizzuto, J.E., Potter, N., Johnson, T.E., Hession, W.C., 2005. The influence of riparian vegetation on stream width, eastern Pennsylvania, USA. GSA Bull. 117, 229–243. https://doi.org/10.1130/B25447.1.

Alvarez, L.V., Schmeeckle, M.W., 2013. Erosion of river sandbars by diurnal stage fluctuations in the colorado river in the marble and grand canyons: full-scale laboratory experiments. River Res. Appl. 29, 839–854. https://doi.org/10.1002/rra.2576.

Andrishak, R., Hicks, F., 2008. Simulating the effects of climate change on the ice regime of the Peace River. Can. J. Civ. Eng. 35, 461–472. https://doi.org/10.1139/L07-129.

Appling, A.P., Bernhardt, E.S., Stanford, J.A., 2014. Floodplain biogeochemical mosaics: a multidimensional view of alluvial soils. J. Geophys. Res. Biogeosci. 119, 1538–1553. https://doi.org/10.1002/2013JG002543.

Ashmore, P., 2013. Morphology and dynamics of braided rivers. In: Wohl, E. (Ed.), Treatise on Geomorphology. Academic Press, San Diego, CA, pp. 290–312.

Ashmore, P., Church, M.A., 2001. The impact of climate change on rivers and river processes in Canada. vol. 555 Geological Survey of Canada, Ottawa, Ontario.

Avis, C.A., Weaver, A.J., Meissner, K.J., 2011. Reduction in areal extent of high-latitude wetlands in response to permafrost thaw. Nat. Geosci. 4, 444–448. https://doi.org/10.1038/ngeo1160.

Bailey, R.G., 1995. Description of the Ecoregions of the United States (Misc. Pub. No. 1391). USDA Forest Service, Washington, DC.

Balcarczyk, K.L., Jones, J.B., Jaffé, R., Maie, N., 2009. Stream dissolved organic matter bioavailability and composition in watersheds underlain with discontinuous permafrost. Biogeochemistry 94, 255–270. https://doi.org/10.1007/s10533-009-9324-x.

Bellmore, J.R., Baxter, C.V., Martens, K., Connolly, P.J., 2012. The floodplain food web mosaic: a study of its importance to salmon and steelhead with implications for their recovery. Ecol. Appl. 23, 189–207. https://doi.org/10.1890/12-0806.1.

Beltaos, S., 2002. Effects of climate on mid-winter ice jams. Hydrol. Process. 16, 789–804. https://doi.org/10.1002/hyp.370.

Beltaos, S., Prowse, T., 2009. River-ice hydrology in a shrinking cryosphere. Hydrol. Process. 23, 122–144. https://doi.org/10.1002/hyp.7165.

Beltaos, S., Prowse, T., Bonsal, B., MacKay, R., Romolo, L., Pietroniro, A., Toth, B., 2006. Climatic effects on ice-jam flooding of the Peace-Athabasca Delta. Hydrol. Process. 20, 4031–4050. https://doi.org/10.1002/hyp.6418.

Bense, V.F., Ferguson, G., Kooi, H., 2009. Evolution of shallow groundwater flow systems in areas of degrading permafrost. Geophys. Res. Lett. 36, L22401. https://doi.org/10.1029/2009GL039225.

Bintanja, R., Selten, F.M., 2014. Future increases in Arctic precipitation linked to local evaporation and sea-ice retreat. Nature 509, 479–482. https://doi.org/10.1038/nature13259.

Blaen, P.J., Hannah, D.M., Brown, L.E., Milner, A.M., 2013. Water temperature dynamics in High Arctic river basins. Hydrol. Process. 27, 2958–2972. https://doi.org/10.1002/hyp.9431.

Bonan, G., Shugart, H.H., 1989. Environmental factors and ecological processes in boreal forests. Annu. Rev. Ecol. Syst. 20, 1–28. https://doi.org/10.1146/annurev.ecolsys.20.1.1.

Boucher, É., Bégin, Y., Arseneault, D., 2009. Impacts of recurring ice jams on channel geometry and geomorphology in a small high-boreal watershed. Geomorphology 108, 273–281. https://doi.org/10.1016/j.geomorph.2009.02.014.

Boucher, E., Bégin, Y., Arseneault, D., Ouarda, T.B.M.J., Church, M., Biron, P.M., Roy, A.G., 2012. Long-term and large-scale river-ice processes in cold-region watersheds. In: Gravel-Bed Rivers. John Wiley & Sons, Ltd., pp. 546–554.

Braudrick, C.A., Grant, G.E., Ishikawa, Y., Ikeda, H., 1997. Dynamics of wood transport in streams: a flume experiment. Earth Surf. Process. Landf. 22, 669–683.

Brooks, A.P., Brierley, G.J., Millar, R.G., 2003. The long-term control of vegetation and woody debris on channel and flood-plain evolution: insights from a paired catchment study in southeastern Australia. Geomorphology 51, 7–29. https://doi.org/10.1016/S0169-555X(02)00323-9.

Brown, J., Ferrians, O., Heginbottom, J.A., Melnikov, E., 2002. Circum-Arctic Map of Permafrost and Ground-Ice Conditions, Version 2. NSIDC: National Snow and Ice Data Center, Boulder, Colorado USA.

Brunskill, G.J., Rosenberg, D.M., Snow, N.B., Vascotto, G.L., Wagemann, R., 1973. Ecological Studies of Aquatic Systems in the Mackenzie-Porcupine Drainages in Relation to Proposed Pipeline and Highway Developments (No. 73–40). Information Canada, Calgary.

Buma, B., Barrett, T.M., 2015. Spatial and topographic trends in forest expansion and biomass change, from regional to local scales. Glob. Change Biol. 21, 3445–3454. https://doi.org/10.1111/gcb.12915.

Burge, L.M., Lapointe, M.F., 2005. Understanding the temporal dynamics of the wandering Renous River, New Brunswick, Canada. Earth Surf. Process. Landf. 30, 1227–1250. https://doi.org/10.1002/esp.1196.

Church, M., 2002. Geomorphic thresholds in riverine landscapes. Freshw. Biol. 47, 541–557. https://doi.org/10.1046/j.1365-2427.2002.00919.x.

Cierjacks, A., Kleinschmit, B., Babinsky, M., Kleinschroth, F., Markert, A., Menzel, M., Ziechmann, U., Schiller, T., Graf, M., Lang, F., 2010. Carbon stocks of soil and vegetation on Danubian floodplains. J. Plant Nutr. Soil Sci. 173, 644–653. https://doi.org/10.1002/jpln.200900209.

Cogley, J.G., McCann, S.B., 1976. An exceptional storm and its effects in the Canadian High Arctic. Arct. Alp. Res. 105–110.

Cole, J.J., Prairie, Y.T., Caraco, N.F., McDowell, W.H., Tranvik, L.J., Striegl, R.G., Duarte, C.M., Kortelainen, P., Downing, J.A., Middelburg, J.J., Melack, J., 2007. Plumbing the global carbon cycle: integrating inland waters into the terrestrial carbon budget. Ecosystems 10, 172–185. https://doi.org/10.1007/s10021-006-9013-8.

Collins, B.D., Montgomery, D.R., Fetherston, K.L., Abbe, T.B., 2012. The floodplain large-wood cycle hypothesis: a mechanism for the physical and biotic structuring of temperate forested alluvial valleys in the North Pacific coastal ecoregion. Geomorphology 139–140, 460–470. https://doi.org/10.1016/j.geomorph.2011.11.011.

Constantine, J.A., Dunne, T., Ahmed, J., Legleiter, C., Lazarus, E.D., 2014. Sediment supply as a driver of river meandering and floodplain evolution in the Amazon Basin. Nat. Geosci. 7, 899–903. https://doi.org/10.1038/ngeo2282.

Costard, F., Dupeyrat, L., Gautier, E., Carey-Gailhardis, E., 2003. Fluvial thermal erosion investigations along a rapidly eroding river bank: application to the Lena River (Central Siberia). Earth Surf. Process. Landf. 28, 1349–1359. https://doi.org/10.1002/esp.592.

Costard, F., Gautier, E., Brunstein, D., Hammadi, J., Fedorov, A., Yang, D., Dupeyrat, L., 2007. Impact of the global warming on the fluvial thermal erosion over the Lena River in Central Siberia. Geophys. Res. Lett. 34, L14501. https://doi.org/10.1029/2007GL030212.

Costard, F., Gautier, E., Fedorov, A., Konstantinov, P., Dupeyrat, L., 2014. An assessment of the erosion potential of the fluvial thermal process during ice breakups of the Lena River (Siberia). Permafr. Periglac. Process. 25, 162–171. https://doi.org/10.1002/ppp.1812.

Danby, R.K., Hik, D.S., 2007. Variability, contingency and rapid change in recent subarctic alpine tree line dynamics. J. Ecol. 95, 352–363. https://doi.org/10.1111/j.1365-2745.2006.01200.x.

Daniels, M.D., Rhoads, B.L., 2004. Effect of large woody debris configuration on three-dimensional flow structure in two low-energy meander bends at varying stages. Water Resour. Res. 40, W11302. https://doi.org/10.1029/2004WR003181.

David, G.C.L., Bledsoe, B.P., Merritt, D.M., Wohl, E., 2009. The impacts of ski slope development on stream channel morphology in the White River National Forest, Colorado, USA. Geomorphology 103, 375–388. https://doi.org/10.1016/j.geomorph.2008.07.003.

Derksen, C., Brown, R., Mudryk, L., Luojus, K., 2015. Terrestrial Snow Cover. in Arctic Report Card 2015.

Devito, K.J., Hill, A.R., Roulet, N., 1996. Groundwater-surface water interactions in headwater forested wetlands of the Canadian Shield. J. Hydrol. 181, 127–147. https://doi.org/10.1016/0022-1694(95)02912-5.

Dingman, S.L., 1973. Effects of permafrost on stream flow characteristics in the discontinuous permafrost zone of Central Alaska. In: North American Contribution to Second International Conference of Permafrost. National Academy of Sciences, Washington, DC, pp. 447–453.

Doetterl, S., Berhe, A.A., Nadeu, E., Wang, Z., Sommer, M., Fiener, P., 2016. Erosion,

deposition and soil carbon: a review of process-level controls, experimental tools and models to address C cycling in dynamic landscapes. Earth-Sci. Rev. 154, 102–122. https://doi.org/10.1016/j.earscirev.2015.12.005.
Doxaran, D., Devred, E., Babin, M., 2015. A 50% increase in the mass of terrestrial particles delivered by the Mackenzie River into the Beaufort Sea (Canadian Arctic Ocean) over the last 10 years. Biogeosciences 12, 3551–3565. https://doi.org/10.5194/bg-12-3551-2015.
Drake, T.W., Wickland, K.P., Spencer, R.G.M., McKnight, D.M., Striegl, R.G., 2015. Ancient low–molecular-weight organic acids in permafrost fuel rapid carbon dioxide production upon thaw. Proc. Natl. Acad. Sci. 112, 13946–13951. https://doi.org/10.1073/pnas.1511705112.
Drury, S.A., Grissom, P.J., 2008. Fire history and fire management implications in the Yukon Flats National Wildlife Refuge, interior Alaska. For. Ecol. Manag. 256, 304–312. https://doi.org/10.1016/j.foreco.2008.04.040.
Dugan, H.A., Lamoureux, S.F., Lafrenière, M.J., Lewis, T., 2009. Hydrological and sediment yield response to summer rainfall in a small high Arctic watershed. Hydrol. Process. 23, 1514–1526. https://doi.org/10.1002/hyp.7285.
Dunne, T., Aalto, R., 2013. Large river floodplains. In: Wohl, E. (Ed.), Treatise on Geomorphology. Academic Press, San Diego, CA, pp. 645–678.
Dunne, T., Mertes, L.A.K., Meade, R.H., Richey, J.E., Forsberg, B.R., 1998. Exchanges of sediment between the flood plain and channel of the Amazon River in Brazil. Geol. Soc. Am. Bull. 110, 450–467.
Dunne, T., Constantine, J.A., Singer, M., 2010. The role of sediment transport and sediment supply in the evolution of river channel and floodplain complexity. Trans. Jpn Geomorphol. Union 31, 155–170.
Dunton, K.H., Schonberg, S.V., Cooper, L.W., 2012. Food web structure of the Alaskan nearshore shelf and estuarine lagoons of the Beaufort Sea. Estuar. Coasts 35, 416–435. http://dx.doi.org.ezproxy2.library.colostate.edu:2048/10.1007/s12237-012-9475-1.
Dupeyrat, L., Costard, F., Randriamazaoro, R., Gailhardis, E., Gautier, E., Fedorov, A., 2011. Effects of ice content on the thermal erosion of permafrost: implications for coastal and fluvial erosion. Permafr. Periglac. Process. 22, 179–187. https://doi.org/10.1002/ppp.722.
Dyke, L.D., Brooks, 2000. The physical environment of the Mackenzie Valley, Northwest Territories: A Base Line for the Assessment of Environmental Change, Geological Survey of Canada, Ottawa, Bulletin vol. 547.
Eaton, B.C., 2006. Bank stability analysis for regime models of vegetated gravel bed rivers. Earth Surf. Process. Landf. 31, 1438–1444. https://doi.org/10.1002/esp.1364.
Epstein, H.E., Raynolds, M.K., Walker, D.A., Bhatt, U.S., Tucker, C.J., Pinzon, J.E., 2012. Dynamics of aboveground phytomass of the circumpolar Arctic tundra during the past three decades. Environ. Res. Lett. 7, 015506. https://doi.org/10.1088/1748-9326/7/1/015506.
Ettema, R., 2006. Ice Effects on Sediment Transport in Rivers. In: Garcia, M. (Ed.), Sedimentation Engineering. American Society of Civil Engineers, pp. 613–648.
Ettema, R., Daly, S.F., 2004. Sediment Transport Under Ice (Cold Regions Research and Engineering Laboratory Report No. TR-04-20). US Army Engineer Research and Development Center, Hanover, NH.
Ettema, R., Kempema, E.W., Church, M., Biron, P.M., Roy, A.G., 2012. River-Ice Effects on Gravel-Bed Channels. In: Gravel-Bed Rivers. John Wiley & Sons, Ltd, pp. 523–540.
Farquharson, L.M., Mann, D.H., Grosse, G., Jones, B.M., Romanovsky, V.E., 2016. Spatial distribution of thermokarst terrain in Arctic Alaska. Geomorphology 273, 116–133. https://doi.org/10.1016/j.geomorph.2016.08.007.
Fischer, L., Kääb, A., Huggel, C., Noetzli, J., 2006. Geology, glacier retreat and permafrost degradation as controlling factors of slope instabilities in a high-mountain rock wall: the Monte Rosa east face. Nat. Hazards Earth Syst. Sci. 6, 761–772. https://doi.org/10.5194/nhess-6-761-2006.
Forbes, A.C., Lamoureux, S.F., 2005. Climatic controls on streamflow and suspended sediment transport in three large middle Arctic catchments, Boothia Peninsula, Nunavut, Canada. Arct. Antarct. Alp. Res. 37, 304–315. https://doi.org/10.1657/1523-0430(2005)037[0304:CCOSAS]2.0.CO;2.
Frey, K.E., McClelland, J.W., 2009. Impacts of permafrost degradation on arctic river biogeochemistry. Hydrol. Process. 23, 169–182. https://doi.org/10.1002/hyp.7196.
Gatto, L.W., 1995. Soil Freeze-Thaw Effects on Bank Erodibility and Stability. DTIC Document.
Gay, G.R., Gay, H.H., Gay, W.H., Martinson, H.A., Meade, R.H., Moody, J.A., 1998. Evolution of cutoffs across meander necks in Powder River, Montana, USA. Earth Surf. Process. Landf. 23, 651–662.
Gooseff, M.N., Balser, A., Bowden, W.B., Jones, J.B., 2009. Effects of hillslope thermokarst in Northern Alaska. EOS Trans. Am. Geophys. Union 90, 29–30. https://doi.org/10.1029/2009EO040001.
Gordeev, V.V., 2006. Fluvial sediment flux to the Arctic Ocean. Geomorphol. Sediment. Source-to-Sink-Fluxes Cold Environ. 80, 94–104. https://doi.org/10.1016/j.geomorph.2005.09.008.
Gorrick, S., Rodríguez, J.F., 2012. Sediment dynamics in a sand bed stream with riparian vegetation. Water Resour. Res. 48. http://dx.doi.org.ezproxy2.library.colostate.edu:2048/10.1029/2011WR011030.
Goulding, H.L., Prowse, T.D., Bonsal, B., 2009. Hydroclimatic controls on the occurrence of break-up and ice-jam flooding in the Mackenzie Delta, NWT, Canada. J. Hydrol. 379, 251–267. https://doi.org/10.1016/j.jhydrol.2009.10.006.
Grant, R.F., 2015. Ecosystem CO2 and CH4 exchange in a mixed tundra and a fen within a hydrologically diverse Arctic landscape: 2. Modeled impacts of climate change. J. Geophys. Res. Biogeosci. 120https://doi.org/10.1002/2014JG002889. 2014JG002889.
Griffin, E.R., Kean, J.W., Vincent, K.R., Smith, J.D., Friedman, J.M., 2005. Modeling effects of bank friction and woody bank vegetation on channel flow and boundary shear stress in the Rio Puerco, New Mexico. J. Geophys. Res. Earth Surf. 110, F04023. https://doi.org/10.1029/2005JF000322.
Güneralp, İ., Rhoads, B.L., 2011. Influence of floodplain erosional heterogeneity on planform complexity of meandering rivers. Geophys. Res. Lett. 38. https://doi.org/10.1029/2011GL048134.
Guo, L., Ping, C.-L., Macdonald, R.W., 2007. Mobilization pathways of organic carbon from permafrost to arctic rivers in a changing climate. Geophys. Res. Lett. 34, L13603. https://doi.org/10.1029/2007GL030689.
Harsch, M.A., Hulme, P.E., McGlone, M.S., Duncan, R.P., 2009. Are treelines advancing? A global meta-analysis of treeline response to climate warming. Ecol. Lett. 12, 1040–1049. https://doi.org/10.1111/j.1461-0248.2009.01355.x.
Hayashi, M., van der Kamp, G., Schmidt, R., 2003. Focused infiltration of snowmelt water in partially frozen soil under small depressions. J. Hydrol. 270, 214–229. https://doi.org/10.1016/S0022-1694(02)00287-1.
Henry, H.A.L., 2008. Climate change and soil freezing dynamics: historical trends and projected changes. Clim. Chang. 87, 421–434. https://doi.org/10.1007/s10584-007-9322-8.
Hession, W.C., Pizzuto, J.E., Johnson, T.E., Horwitz, R.J., 2003. Influence of bank vegetation on channel morphology in rural and urban watersheds. Geology 31, 147–150. https://doi.org/10.1130/0091-7613(2003)031<0147:IOBVOC>2.0.CO;2.
Hicks, F.E., 1993. Ice as the geomorphologic agent in an anastomosing river system. In: Proceedings of the 7th Workshop on the Hydraulics of Ice Covered Rivers, Saskatoon, Sask. NHRI Symposium Series, pp. 3–20.
Hicks, F., 2009. An overview of river ice problems: CRIPE07 guest editorial. Cold Reg. Sci. Technol. 55, 175–185. https://doi.org/10.1016/j.coldregions.2008.09.006.
Hicks, F., Beltaos, S., 2008. River ice. In: Woo, M. (Ed.), Cold Region Atmospheric and Hydrologic Studies. The Mackenzie GEWEX Experience. Springer, Berlin Heidelberg, pp. 281–305. https://doi.org/10.1007/978-3-540-75136-6_15.
Hilton, R.G., Galy, V., Gaillardet, J., Dellinger, M., Bryant, C., O'Regan, M., Gröcke, D.R., Coxall, H., Bouchez, J., Calmels, D., 2015. Erosion of organic carbon in the Arctic as a geological carbon dioxide sink. Nature 524, 84–87. https://doi.org/10.1038/nature14653.
Hinzman, L.D., Deal, C.J., McGuire, A.D., Mernild, S.H., Polyakov, I.V., Walsh, J.E., 2013. Trajectory of the Arctic as an integrated system. Ecol. Appl. 23, 1837–1868. https://doi.org/10.1890/11-1498.1.
Hjulstrom, F., 1935. Studies of the morphological activity of rivers as illustrated by the River Fyris. Bull. Geol. Inst. Upsalsa 25, 221–527.
Holland, M.M., Finnis, J., Serreze, M.C., 2006. Simulated Arctic ocean freshwater budgets in the twentieth and twenty-first centuries. J. Clim. 19, 6221–6242. https://doi.org/10.1175/JCLI3967.1.
Holmes, R.M., McClelland, J.W., Peterson, B.J., Shiklomanov, I.A., Shiklomanov, A.I., Zhulidov, A.V., Gordeev, V.V., Bobrovitskaya, N.N., 2002. A circumpolar perspective on fluvial sediment flux to the Arctic Ocean. Glob. Biogeochem. Cycles 16 (1098). https://doi.org/10.1029/2001GB001849.
Holmes, R.M., McClelland, J.W., Peterson, B.J., Tank, S.E., Bulygina, E., Eglinton, T.I., Gordeev, V.V., Gurtovaya, T.Y., Raymond, P.A., Repeta, D.J., Staples, R., Striegl, R.G., Zhulidov, A.V., Zimov, S.A., 2012. Seasonal and annual fluxes of nutrients and organic matter from large rivers to the Arctic ocean and surrounding seas. Estuar. Coasts 35, 369–382. https://doi.org/10.1007/s12237-011-9386-6.
Holmes, R.M., Coe, M.T., Fiske, G.J., Gurtovaya, T., McClelland, J.W., Shiklomanov, A.I., Spencer, R.G., Tank, S.E., Zhulidov, A.V., Goldman, C.R., Kumagai, M., Robarts, R.D., 2013. Climate change impacts on the hydrology and biogeochemistry of Arctic rivers. In: Climatic Change and Global Warming of Inland Waters: Impacts and Mitigation for Ecosystems and Societies. John Wiley & Sons Ltd, pp. 3–26.
Hugelius, G., Strauss, J., Zubrzycki, S., Harden, J.W., Schuur, E.A.G., Ping, C.-L., Schirrmeister, L., Grosse, G., Michaelson, G.J., Koven, C.D., O'Donnell, J.A., Elberling, B., Mishra, U., Camill, P., Yu, Z., Palmtag, J., Kuhry, P., 2014. Estimated stocks of circumpolar permafrost carbon with quantified uncertainty ranges and identified data gaps. Biogeosciences 11, 6573–6593. https://doi.org/10.5194/bg-11-6573-2014.
Huisink, M., de Moor, J.J.W., Kasse, C., Virtanen, T., 2002. Factors influencing periglacial fluvial morphology in the northern European Russian tundra and taiga. Earth Surf. Process. Landf. 27, 1223–1235. https://doi.org/10.1002/esp.422.
IPCC, 2014. Climate Change 2014: Synthesis Report. Contribution of Working Groups I, II, III to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change. IPCC, Geneva, Switzerland.
Järvelä, J., 2004. Determination of flow resistance caused by non-submerged woody vegetation. Int. J. River Basin Manag. 2, 61–70.
Jeffries, R., Darby, S.E., Sear, D.A., 2003. The influence of vegetation and organic debris on flood-plain sediment dynamics: case study of a low-order stream in the New Forest, England. Geomorphology 51, 61–80. https://doi.org/10.1016/S0169-555X(02)00325-2.
Jobbágy, E.G., Jackson, R.B., 2000. The vertical distribution of soil organic carbon and its relation to climate and vegetation. Ecol. Appl. 10, 423–436.
Johnson, S.L., Swanson, F.J., Grant, G.E., Wondzell, S.M., 2000. Riparian forest disturbances by a mountain flood - the influence of floated wood. Hydrol. Process. 14, 3031–3050.
Jorgenson, M.T., Racine, C.H., Walters, J.C., Osterkamp, T.E., 2001. Permafrost degradation and ecological changes associated with a warming climate in Central Alaska. Clim. Chang. 48, 551–579. https://doi.org/10.1023/A:1005667424292.
Jorgenson, M.T., Shur, Y.L., Pullman, E.R., 2006. Abrupt increase in permafrost degradation in Arctic Alaska. Geophys. Res. Lett. 33, L02503. https://doi.org/10.1029/2005GL024960.
Jorgenson, M.T., Harden, J., Kanevskiy, M., O'Donnell, J., Wickland, K., Ewing, Stephanie, Manies, K., Zhuang, Q., Shur, Y., Striegl, R., Koch, J., 2013. Reorganization of vegetation, hydrology and soil carbon after permafrost degradation

across heterogeneous boreal landscapes. Environ. Res. Lett. 8, 035017. https://doi.org/10.1088/1748-9326/8/3/035017.
Jorgenson, J., Jorgenson, M., Boldenow, M., Orndahl, K., Jorgenson, J.C., Jorgenson, M.T., Boldenow, M.L., Orndahl, K.M., 2018. Landscape change detected over a half century in the Arctic National Wildlife refuge using high-resolution aerial imagery. Remote Sens. 10, 1305. https://doi.org/10.3390/rs10081305.
Junk, W.J., Bayley, P.B., Sparks, R.E., 1989. The Flood Pulse Concept in River-Floodplain Systems. In: Proceedings of the International Large River Symposium, Can. Spec. Publ. Fish. Aquat. Sci. vol. 106. D.P. Dodge, pp. 110–127.
Kane, D.L., Hinzman, L.D., Gieck, R.E., McNamara, J.P., Youcha, E.K., Oatley, J.A., 2008. Contrasting extreme runoff events in areas of continuous permafrost, Arctic Alaska. Hydrol. Res. 39, 287–298. https://doi.org/10.2166/nh.2008.005.
Kasischke, E.S., Turetsky, M.R., 2006. Recent changes in the fire regime across the north American boreal region—Spatial and temporal patterns of burning across Canada and Alaska. Geophys. Res. Lett. 33. https://doi.org/10.1029/2006GL025677.
Kimiaghalam, N., Goharrokhi, M., Clark, S.P., Ahmari, H., 2015. A comprehensive fluvial geomorphology study of riverbank erosion on the Red River in Winnipeg, Manitoba, Canada. J. Hydrol. 529, 1488–1498. https://doi.org/10.1016/j.jhydrol.2015.08.033.
Kirtman, B., Power, S.B., Adedoyin, A.J., Boer, G.J., Bojariu, R., Camilloni, I., Doblas-Reyes, F., Fiore, A.M., Kimoto, M., Meehl, G., 2013. Near-term Climate Change: Projections and Predictability.
Koenigk, T., Brodeau, L., Graversen, R.G., Karlsson, J., Svensson, G., Tjernström, M., Willén, U., Wyser, K., 2013. Arctic climate change in 21st century CMIP5 simulations with EC-Earth. Clim. Dyn. 40, 2719–2743. https://doi.org/10.1007/s00382-012-1505-y.
Kokelj, S.V., Lacelle, D., Lantz, T.C., Tunnicliffe, J., Malone, L., Clark, I.D., Chin, K.S., 2013. Thawing of massive ground ice in mega slumps drives increases in stream sediment and solute flux across a range of watershed scales. J. Geophys. Res. Earth Surf. 118, 681–692. https://doi.org/10.1002/jgrf.20063.
Koven, C.D., Ringeval, B., Friedlingstein, P., Ciais, P., Cadule, P., Khvorostyanov, D., Krinner, G., Tarnocai, C., 2011. Permafrost carbon-climate feedbacks accelerate global warming. Proc. Natl. Acad. Sci. 108, 14769–14774. https://doi.org/10.1073/pnas.1103910108.
Krasting, J.P., Broccoli, A.J., Dixon, K.W., Lanzante, J.R., 2013. Future changes in northern hemisphere snowfall. J. Clim. 26, 7813–7828. https://doi.org/10.1175/JCLI-D-12-00832.1.
Kuhnle, R.A., 2013. Suspended load. In: Wohl, E. (Ed.), Treatise on Geomorphology. Academic Press, San Diego, CA, pp. 124–136.
Lamoureux, S.F., Lafrenière, M.J., 2009. Fluvial impact of extensive active layer detachments, Cape Bounty, Melville Island, Canada. Arct. Antarct. Alp. Res. 41, 59–68. https://doi.org/10.1657/1523-0430-41.1.59.
Lawler, D., Carling, P.A., Petts, G.E., 1992. Process dominance in bank erosion systems. In: Lowland Floodplain Rivers: Geomorphological Perspectives. John Wiley & Sons Ltd, pp. 117–143.
Lawler, D.M., 1993. Needle ice processes and sediment mobilization on river banks: the River Ilston, West Glamorgan, UK. J. Hydrol. 150, 81–114. https://doi.org/10.1016/0022-1694(93)90157-5.
Lawrence, D.M., Slater, A.G., 2005. A projection of severe near-surface permafrost degradation during the 21st century. Geophys. Res. Lett. 32. https://doi.org/10.1029/2005GL025080.
Le Lay, T., Piegay, H., Moulin, B., 2013. Wood entrance, deposition, transfer, and effects on fluvial forms and processes: Problem statements and challenging issue. In: Butler, D.R., Hupp, C.R. (Eds.), Treatise on Geomorphology. Academic Press, San Diego, CA, pp. 20–36.
Leopold, L.B., 1973. River channel change with time: an example: address as retiring president of the Geological Society of America, Minneapolis, Minnesota, November 1972. Geol. Soc. Am. Bull. 84, 1845–1860.
Leopold, L.B., Wolman, M.G., Miller, J.P., 1964. Fluvial Processes in Geomorphology, Series of Books in Geology. W. H. Freeman, San Francisco.
Lewin, J., Ashworth, P.J., 2014. The negative relief of large river floodplains. Earth-Sci. Rev. 129, 1–23. https://doi.org/10.1016/j.earscirev.2013.10.014.
Lewis, T., Lamoureux, S.F., 2010. Twenty-first century discharge and sediment yield predictions in a small high Arctic watershed. Glob. Planet. Change 71, 27–41. https://doi.org/10.1016/j.gloplacha.2009.12.006.
Lininger, K.B., Latrubesse, E.M., 2016. Flooding hydrology and peak discharge attenuation along the middle Araguaia River in Central Brazil. CATENA 143, 90–101. https://doi.org/10.1016/j.catena.2016.03.043.
Lininger, K.B., Wohl, E., Sutfin, N.A., Rose, J.R., 2017. Floodplain downed wood volumes: a comparison across three biomes. Earth Surf. Process. Landf. 42, 1248–1261. https://doi.org/10.1002/esp.4072.
Lininger, K.B., Wohl, E., Rose, J.R., 2018. Geomorphic Controls on Floodplain Soil Organic Carbon in the Yukon Flats, Interior Alaska, from Reach to River Basin Scales. Water Resour. Res. 1934–1951. https://doi.org/10.1002/2017WR022042.
Lininger, K.B., Wohl, E., Rose, J.R., Leisz, S.J., 2019. Significant floodplain soil organic carbon storage along a large high latitude river and its tributaries. Geophys. Res. Lett. 46. https://doi.org/10.1029/2018GL080996.
Lloyd, A.H., Yoshikawa, K., Fastie, C.L., Hinzman, L., Fraver, M., 2003. Effects of permafrost degradation on woody vegetation at arctic treeline on the Seward Peninsula, Alaska. Permafr. Periglac. Process. 14, 93–101. https://doi.org/10.1002/ppp.446.
Loisel, J., van Bellen, S., Pelletier, L., Talbot, J., Hugelius, G., Karran, D., Yu, Z., Nichols, J., Holmquist, J., 2017. Insights and issues with estimating northern peatland carbon stocks and fluxes since the last Glacial Maximum. Earth-Sci. Rev. 165, 59–80. https://doi.org/10.1016/j.earscirev.2016.12.001.
Mackay, D.K., Fogarasi, S., Spitzer, M., 1973. Documentation of an extreme summer storm in the Mackenzie Mountains, N.W.T. (no. 73–3). In: Hydrologic Aspects of Northern Pipeline Development. Glacial Division, Water Resources Branch, Department of the Environment.
Manga, M., Kirchner, J.W., 2000. Stress partitioning in streams by large woody debris. Water Resour. Res. 36, 2373–2379. https://doi.org/10.1029/2000WR900153.
Mann, D.H., Groves, P., Reanier, R.E., Kunz, M.L., 2010. Floodplains, permafrost, cottonwood trees, and peat: what happened the last time climate warmed suddenly in arctic Alaska? Quat. Sci. Rev. 29, 3812–3830. https://doi.org/10.1016/j.quascirev.2010.09.002.
Martin, D.J., Benda, L.E., 2001. Patterns of instream wood recruitment and transport at the watershed scale. Trans. Am. Fish. Soc. 130, 940–958. https://doi.org/10.1577/1548-8659(2001)130<0940:POIWRA>2.0.CO;2.
Martín-Vide, J.P., Amarilla, M., Zárate, F.J., 2014. Collapse of the Pilcomayo River. In: Geomorphology, Discontinuities in Fluvial Systems 205, pp. 155–163. https://doi.org/10.1016/j.geomorph.2012.12.007.
McClelland, J.W., Holmes, R.M., Dunton, K.H., Macdonald, R.W., 2012. The Arctic ocean estuary. Estuar. Coasts 35, 353–368. http://dx.doi.org.ezproxy2.library.colostate.edu:2048/10.1007/s12237-010-9357-3.
McClelland, J.W., Holmes, R.M., Peterson, B.J., Raymond, P.A., Striegl, R.G., Zhulidov, A.V., Zimov, S.A., Zimov, N., Tank, S.E., Spencer, R.G.M., Staples, R., Gurtovaya, T.Y., Griffin, C.G., 2016. Particulate organic carbon and nitrogen export from major Arctic rivers. Glob. Biogeochem. Cycles 30, 629–643. https://doi.org/10.1002/2015GB005351.
McKenney, R., Jacobson, R.B., Wertheimer, R.C., 1995. Woody vegetation and channel morphogenesis in low-gradient, gravel-bed streams in the Ozark Plateaus, Missouri and Arkansas. In: Hupp, C.R., Osterkamp, W.R., Howard, A.D. (Eds.), Biogeomorphology, Terrestrial and Freshwater Systems. Elsevier, Amsterdam, pp. 175–198. https://doi.org/10.1016/B978-0-444-81867-6.50016-2.
Meade, R.H., 2007. Transcontinental moving and storage: The Orinoco and Amazon Rivers transfer the Andes to the Atlantic. In: Large Rivers Geomorphol. Manag. pp. 45–64.
Merritt, D.M., 2013. Reciprocal relations between riparian vegetation, fluvial landforms, and channel processes. In: Wohl, E. (Ed.), Treatise on Geomorphology. Academic Press, San Diego, CA, pp. 220–243.
Mertes, L.A.K., 1997. Documentation and significance of the perirheic zone on inundated floodplains. Water Resour. Res. 33, 1749–1762. https://doi.org/10.1029/97WR00658.
Mertes, L.A.K., Dunne, T., 2007. Effects of tectonism, climate change, and sea-level change on the form and behaviour of the modern Amazon river and its floodplain. In: Gupta, A. (Ed.), Large Rivers. John Wiley & Sons, Ltd, pp. 115–144.
Mertes, L.A.K., Dunne, T., Martinelli, L.A., 1996. Channel-floodplain geomorphology along the Solimões-Amazon River, Brazil. Geol. Soc. Am. Bull. 108, 1089–1107. https://doi.org/10.1130/0016-7606(1996)108<1089:CFGATS>2.3.CO;2.
Min, S.-K., Zhang, X., Zwiers, F., 2008. Human-induced Arctic moistening. Science 320, 518–520. https://doi.org/10.1126/science.1153468.
Minsley, B.J., Abraham, J.D., Smith, B.D., Cannia, J.C., Voss, C.I., Jorgenson, M.T., Walvoord, M.A., Wylie, B.K., Anderson, L., Ball, L.B., Deszcz-Pan, M., Wellman, T.P., Ager, T.A., 2012. Airborne electromagnetic imaging of discontinuous permafrost. Geophys. Res. Lett. 39https://doi.org/10.1029/2011GL050079. n/a-n/a.
Montgomery, D.R., 2002. Valley formation by fluvial and glacial erosion. Geology 30, 1047–1050. https://doi.org/10.1130/0091-7613(2002)030<1047:VFBFAG>2.0.CO;2.
Morse, P.D., Wolfe, S.A., 2015. Geological and meteorological controls on icing (aufeis) dynamics (1985 to 2014) in subarctic Canada: icing dynamics in Subarctic Canada. J. Geophys. Res. Earth Surf. 120, 1670–1686. https://doi.org/10.1002/2015JF003534.
Murton, J.B., Belshaw, R.K., 2011. A conceptual model of valley incision, planation and terrace formation during cold and arid permafrost conditions of Pleistocene southern England. Quat. Res. In: Special Issue on Periglacial Processes, Landforms, and Environments in Honor of Link Washburn. vol. 75. pp. 385–394. https://doi.org/10.1016/j.yqres.2010.10.002.
Myers-Smith, I.H., Harden, J.W., Wilmking, M., Fuller, C.C., Mcguire, A.D., Chapin III, F.S., 2008. Wetland succession in a permafrost collapse: interactions between fire and thermokarst. Biogeosci. Discuss. 5, 1273–1286.
Nanson, G.C., Croke, J.C., 1992. A genetic classification of floodplains. Geomorphology 4, 459–486.
Nanson, G.C., Knighton, A.D., 1996. Anabranching rivers: their cause, character and classification. Earth Surf. Process. Landf. 21, 217–239. https://doi.org/10.1002/(SICI)1096-9837(199603)21:3<217::AID-ESP611>3.0.CO;2-U.
Nelson, F.E., Anisimov, O.A., Shiklomanov, N.I., 2002. Climate change and hazard zonation in the circum-arctic permafrost regions. Nat. Hazards 26, 203–225. https://doi.org/10.1023/A:1015612918401.
Nilsson, C., Jansson, R., Kuglerová, L., Lind, L., Ström, L., 2012. Boreal riparian vegetation under climate change. Ecosystems 16, 401–410. https://doi.org/10.1007/s10021-012-9622-3.
Nilsson, C., Polvi, L.E., Lind, L., 2015. Extreme events in streams and rivers in arctic and subarctic regions in an uncertain future. Freshw. Biol. https://doi.org/10.1111/fwb.12477.
O'Connor, J.E., Grant, G.E., Costa, J.E., 2002. The geology and geography of floods. In: House, P.K., Webb, R.H., Baker, V.R., Levish, D.R. (Eds.), Ancient Floods, Modern Hazards, Water Science and Application. American Geophysical Union Press, Washington, DC, pp. 359–385.
O'Connor, J.E., Jones, M.A., Haluska, T.L., 2003. Flood plain and channel dynamics of the Quinault and Queets Rivers, Washington, USA. Geomorphology 51, 31–59. https://doi.org/10.1016/S0169-555X(02)00324-0.
O'Donnell, J.A., Harden, J.W., McGuire, A.D., Kanevskiy, M.Z., Jorgenson, M.T., Xu, X., 2011. The effect of fire and permafrost interactions on soil carbon accumulation in an upland black spruce ecosystem of interior Alaska: implications for post-thaw carbon loss. Glob. Change Biol. 17, 1461–1474. https://doi.org/10.1111/j.1365-2486.2010.

02358.x.
Olefeldt, D., Goswami, S., Grosse, G., Hayes, D., Hugelius, G., Kuhry, P., McGuire, A.D., Romanovsky, V.E., Sannel, A.B.K., Schuur, E.A.G., Turetsky, M.R., 2016. Circumpolar distribution and carbon storage of thermokarst landscapes. Nat. Commun. 7 (13043). https://doi.org/10.1038/ncomms13043.
Osterkamp, T.E., 2005. The recent warming of permafrost in Alaska. Glob. Planet. Change 49, 187–202. https://doi.org/10.1016/j.gloplacha.2005.09.001.
Osterkamp, T.E., Jorgenson, M.T., Schuur, E.A.G., Shur, Y.L., Kanevskiy, M.Z., Vogel, J.G., Tumskoy, V.E., 2009. Physical and ecological changes associated with warming permafrost and thermokarst in Interior Alaska. Permafr. Periglac. Process. 20, 235–256. https://doi.org/10.1002/ppp.656.
Ott, R.A., Lee, M.A., Putman, W.E., Mason, O.K., Worum, G.T., Burns, D.N., 2001. Bank erosion and large woody debris recruitment along the Tanana River, interior Alaska. In: Alaska Department of Natural Resources Division of Forestry and Tanana Chiefs Conference, Inc. Forestry Program.
Overeem, I., Syvitski, J.P.M., 2010. Shifting discharge peaks in Arctic rivers, 1977–2007. Geogr. Ann. Ser. Phys. Geogr. 92, 285–296. https://doi.org/10.1111/j.1468-0459.2010.00395.x.
Park, H., Walsh, J., Fedorov, A.N., Sherstiukov, A.B., Iijima, Y., Ohata, T., 2013. The influence of climate and hydrological variables on opposite anomaly in active-layer thickness between Eurasian and north American watersheds. Cryosphere 7, 631–645. https://doi.org/10.5194/tc-7-631-2013.
Pastick, N.J., Rigge, M., Wylie, B.K., Jorgenson, M.T., Rose, J.R., Johnson, K.D., Ji, L., 2014. Distribution and landscape controls of organic layer thickness and carbon within the Alaskan Yukon River Basin. Geoderma 230–231, 79–94. https://doi.org/10.1016/j.geoderma.2014.04.008.
Pastick, N.J., Jorgenson, M.T., Wylie, B.K., Nield, S.J., Johnson, K.D., Finley, A.O., 2015. Distribution of near-surface permafrost in Alaska: estimates of present and future conditions. Remote Sens. Environ. 168, 301–315. https://doi.org/10.1016/j.rse.2015.07.019.
Pavelsky, T.M., Zarnetske, J.P., 2017. Rapid decline in river icings detected in Arctic Alaska: implications for a changing hydrologic cycle and river ecosystems: river Icing declines in Arctic Alaska. Geophys. Res. Lett. 44, 3228–3235. https://doi.org/10.1002/2016GL072397.
Peterson, B.J., Holmes, R.M., McClelland, J.W., Vörösmarty, C.J., Lammers, R.B., Shiklomanov, A.I., Shiklomanov, I.A., Rahmstorf, S., 2002. Increasing river discharge to the Arctic ocean. Science 298, 2171–2173. https://doi.org/10.1126/science.1077445.
Petrone, K.C., Jones, J.B., Hinzman, L.D., Boone, R.D., 2006. Seasonal export of carbon, nitrogen, and major solutes from Alaskan catchments with discontinuous permafrost. J. Geophys. Res. Biogeosci. 111, G02020. https://doi.org/10.1029/2005JG000055.
Pfirman, S.L., Eicken, H., Bauch, D., Weeks, W.F., 1995. The potential transport of pollutants by Arctic sea ice. Sci. Total Environ. 159, 129–146. https://doi.org/10.1016/0048-9697(95)04174-Y.
Phillips, J.D., 2003. Sources of nonlinearity and complexity in geomorphic systems. Prog. Phys. Geogr. Earth Environ. 27, 1–23. https://doi.org/10.1191/0309133303pp340ra.
Piegay, H., 1993. Nature, mass and preferential sites of coarse woody debris deposits in the lower Ain valley (Mollon reach), France. Regul. Rivers Res. Manag. 8, 359–372. https://doi.org/10.1002/rrr.3450080406.
Pinay, G., Fabre, A., Vervier, P., Gazelle, F., 1992. Control of C,N,P distribution in soils of riparian forests. Landsc. Ecol. 6, 121–132. https://doi.org/10.1007/BF00130025.
Pizzuto, J.E., 2009. An empirical model of event scale cohesive bank profile evolution. Earth Surf. Process. Landf. 34, 1234–1244. https://doi.org/10.1002/esp.1808.
Pollen, N., Simon, A., 2005. Estimating the mechanical effects of riparian vegetation on stream bank stability using a fiber bundle model. Water Resour. Res. 41, W07025. https://doi.org/10.1029/2004WR003801.
Polvi, L.E., Wohl, E., Merritt, D.M., 2014. Modeling the functional influence of vegetation type on streambank cohesion. Earth Surf. Process. Landf. 39, 1245–1258. https://doi.org/10.1002/esp.3577.
Prowse, T.D., 1994. Environmental significance of ice to streamflow in cold regions. Freshw. Biol. 32, 241–259. https://doi.org/10.1111/j.1365-2427.1994.tb01124.x.
Prowse, T.D., Beltaos, S., 2002. Climatic control of river-ice hydrology: a review. Hydrol. Process. 16, 805–822. https://doi.org/10.1002/hyp.369.
Prowse, T.D., Carter, T., 2002. Significance of ice-induced storage to spring runoff: a case study of the Mackenzie River. Hydrol. Process. 16, 779–788. https://doi.org/10.1002/hyp.371.
Prowse, T.D., Culp, J.M., 2003. Ice breakup: a neglected factor in river ecology. Can. J. Civ. Eng. 30, 128.
Prowse, T.D., Ferrick, M.G., 2002. Hydrology of ice-covered rivers and lakes: scoping the subject. Hydrol. Process. 16, 759–762. https://doi.org/10.1002/hyp.373.
Prowse, T.D., Wrona, F.J., Reist, J.D., Gibson, J.J., Hobbie, J.E., Lévesque, L.M.J., Vincent, W.F., 2006. Climate change effects on hydroecology of Arctic freshwater ecosystems. AMBIO J. Hum. Environ. 35, 347–358. https://doi.org/10.1579/0044-7447(2006)35[347:CCEOHO]2.0.CO;2.
Prowse, T., Alfredsen, K., Beltaos, S., Bonsal, B., Duguay, C., Korhola, A., McNamara, J., Pienitz, R., Vincent, W.F., Vuglinsky, V., Weyhenmeyer, G.A., 2012. Past and future changes in Arctic Lake and River ice. AMBIO 40, 53–62. https://doi.org/10.1007/s13280-011-0216-7.
Quinton, W.L., Hayashi, M., Chasmer, L.E., 2009. Peatland hydrology of discontinuous permafrost in the northwest territories: overview and synthesis. Can. Water Resour. J. Rev. Can. Ressour. Hydr. 34, 311–328. https://doi.org/10.4296/cwrj3404311.
Quinton, W.L., Hayashi, M., Chasmer, L.E., 2011. Permafrost-thaw-induced land-cover change in the Canadian subarctic: implications for water resources. Hydrol. Process. 25, 152–158. https://doi.org/10.1002/hyp.7894.
Randriamazaoro, R., Dupeyrat, L., Costard, F., Carey Gailhardis, E., 2007. Fluvial thermal erosion: heat balance integral method. Earth Surf. Process. Landf. 32, 1828–1840. https://doi.org/10.1002/esp.1489.
Rawlins, M.A., Steele, M., Holland, M.M., Adam, J.C., Cherry, J.E., Francis, J.A., Groisman, P.Y., Hinzman, L.D., Huntington, T.G., Kane, D.L., Kimball, J.S., Kwok, R., Lammers, R.B., Lee, C.M., Lettenmaier, D.P., McDonald, K.C., Podest, E., Pundsack, J.W., Rudels, B., Serreze, M.C., Shiklomanov, A., Skagseth, Ø., Troy, T.J., Vörösmarty, C.J., Wensnahan, M., Wood, E.F., Woodgate, R., Yang, D., Zhang, K., Zhang, T., 2010. Analysis of the Arctic system for freshwater cycle intensification: observations and expectations. J. Clim. 23, 5715–5737. https://doi.org/10.1175/2010JCLI3421.1.
Raymond, P.A., McClelland, J.W., Holmes, R.M., Zhulidov, A.V., Mull, K., Peterson, B.J., Striegl, R.G., Aiken, G.R., Gurtovaya, T.Y., 2007. Flux and age of dissolved organic carbon exported to the Arctic Ocean: a carbon isotopic study of the five largest arctic rivers. Glob. Biogeochem. Cycles 21, GB4011. https://doi.org/10.1029/2007GB002934.
Regnier, P., Friedlingstein, P., Ciais, P., Mackenzie, F.T., Gruber, N., Janssens, I.A., Laruelle, G.G., Lauerwald, R., Luyssaert, S., Andersson, A.J., Arndt, S., Arnosti, C., Borges, A.V., Dale, A.W., Gallego-Sala, A., Goddéris, Y., Goossens, N., Hartmann, J., Heinze, C., Ilyina, T., Joos, F., LaRowe, D.E., Leifeld, J., Meysman, F.J.R., Munhoven, G., Raymond, P.A., Spahni, R., Suntharalingam, P., Thullner, M., 2013. Anthropogenic perturbation of the carbon fluxes from land to ocean. Nat. Geosci. 6, 597–607. https://doi.org/10.1038/ngeo1830.
Rennermalm, A.K., Wood, E.F., Troy, T.J., 2010. Observed changes in pan-arctic cold-season minimum monthly river discharge. Clim. Dyn. 35, 923–939. http://dx.doi.org.ezproxy2.library.colostate.edu:2048/10.1007/s00382-009-0730-5.
Rinaldi, M., Darby, S.E., 2007. 9 Modelling river-bank-erosion processes and mass failure mechanisms: Progress towards fully coupled simulations. In: Helmut Habersack, H.P. (Ed.), Developments in Earth Surface Processes, Gravel-Bed Rivers VI: From Process Understanding to River Restoration. Elsevier, pp. 213–239.
Rinaldi, M., Casagli, N., Dapporto, S., Gargini, A., 2004. Monitoring and modelling of pore water pressure changes and riverbank stability during flow events. Earth Surf. Process. Landf. 29, 237–254. https://doi.org/10.1002/esp.1042.
Rinaldi, M., Mengoni, B., Luppi, L., Darby, S.E., Mosselman, E., 2008. Numerical simulation of hydrodynamics and bank erosion in a river bend. Water Resour. Res. 44, W09428. https://doi.org/10.1029/2008WR007008.
Rokaya, P., Budhathoki, S., Lindenschmidt, K.-E., 2018. Trends in the timing and magnitude of ice-jam floods in Canada. Sci. Rep. 8 (5834). https://doi.org/10.1038/s41598-018-24057-z.
Romanovsky, V.E., Smith, S.L., Christiansen, H.H., Shiklomanov, N.I., Streletskiy, D.A., Drozdov, D.S., Oberman, N.G., Kholodov, A.L., Marchenko, S.S., 2013. Permafrost. Arctic Report Card 2013.
Rowland, J.C., Jones, C.E., Altmann, G., Bryan, R., Crosby, B.T., Hinzman, L.D., Kane, D.L., Lawrence, D.M., Mancino, A., Marsh, P., McNamara, J.P., Romanovsky, V.E., Toniolo, H., Travis, B.J., Trochim, E., Wilson, C.J., Geernaert, G.L., 2010. Arctic landscapes in transition: responses to Thawing Permafrost. EOS Trans. Am. Geophys. Union 91, 229–230. https://doi.org/10.1029/2010EO260001.
Schreiner, K.M., Bianchi, T.S., Rosenheim, B.E., 2014. Evidence for permafrost thaw and transport from an Alaskan North Slope watershed. Geophys. Res. Lett. 41, 3117–3126. https://doi.org/10.1002/2014GL059514.
Schumm, S.A., 1979. Geomorphic thresholds: the concept and its applications. Trans. Inst. Br. Geogr. 4, 485–515. https://doi.org/10.2307/622211.
Schumm, S.A., 1993. River response to baselevel change: implications for sequence stratigraphy. J. Geol. 101, 279–294.
Schuur, E.A.G., Bockheim, J., Canadell, J.G., Euskirchen, E., Field, C.B., Goryachkin, S.V., Hagemann, S., Kuhry, P., Lafleur, P.M., Lee, H., Mazhitova, G., Nelson, F.E., Rinke, A., Romanovsky, V.E., Shiklomanov, N., Tarnocai, C., Venevsky, S., Vogel, J.G., Zimov, S.A., 2008. Vulnerability of permafrost carbon to climate change: implications for the global carbon cycle. BioScience 58, 701–714. https://doi.org/10.1641/B580807.
Schuur, E.A.G., McGuire, A.D., Schädel, C., Grosse, G., Harden, J.W., Hayes, D.J., Hugelius, G., Koven, C.D., Kuhry, P., Lawrence, D.M., Natali, S.M., Olefeldt, D., Romanovsky, V.E., Schaefer, K., Turetsky, M.R., Treat, C.C., Vonk, J.E., 2015. Climate change and the permafrost carbon feedback. Nature 520, 171–179. https://doi.org/10.1038/nature14338.
Scott, K.M., 1978. Effects of Permafrost on Stream Channel Behavior in Arctic Alaska (no. 1068). U.S. Geological Survey, Washington, DC.
Semmens, K.A., Ramage, J.M., 2013. Recent changes in spring snowmelt timing in the Yukon River basin detected by passive microwave satellite data. Cryosphere 7, 905–916. https://doi.org/10.5194/tc-7-905-2013.
Serreze, M.C., Walsh, J.E., Chapin, F.S., Osterkamp, T., al, et, 2000. Observational evidence of recent change in the northern high-latitude environment. Clim. Chang. 46, 159–207. https://doi.org/10.1023/A:1005504031923.
Shelef, E., 21 February, 2019. Personal Communication on the Percent of Floodplain Area in Circumpolar Permafrost Regions.
Sholtes, J., Doyle, M., 2011. Effect of channel restoration on flood wave attenuation. J. Hydraul. Eng. 137, 196–208. https://doi.org/10.1061/(ASCE)HY.1943-7900.0000294.
Simon, A., Curini, A., Darby, S.E., Langendoen, E.J., 2000. Bank and near-bank processes in an incised channel. Geomorphology 35, 193–217. https://doi.org/10.1016/S0169-555X(00)00036-2.
Simons, D.B., Richardson, E.V., Haushild, W.L., 1963. Some Effects of Fine Sediment on Flow Phenomena (USGS Numbered Series no. 1498- G). Water Supply Paper. U.S. Geological Survey.
Slater, A.G., Lawrence, D.M., 2013. Diagnosing present and future permafrost from climate models. J. Clim. 26, 5608–5623. https://doi.org/10.1175/JCLI-D-12-00341.1.
Smith, D.G., 1979. Effects of channel enlargement by river ice processes on bankfull

discharge in Alberta, Canada. Water Resour. Res. 15, 469–475. https://doi.org/10.1029/WR015i002p00469.
Smith, D.G., Pearce, C.M., 2002. Ice jam-caused fluvial gullies and scour holes on northern river flood plains. Geomorphology 42, 85–95. https://doi.org/10.1016/S0169-555X(01)00076-9.
Spence, C., Woo, M., 2003. Hydrology of subarctic Canadian shield: soil-filled valleys. J. Hydrol. 279, 151–166. https://doi.org/10.1016/S0022-1694(03)00175-6.
Spencer, R.G.M., Mann, P.J., Dittmar, T., Eglinton, T.I., McIntyre, C., Holmes, R.M., Zimov, N., Stubbins, A., 2015. Detecting the signature of permafrost thaw in Arctic rivers. Geophys. Res. Lett. 42, 2830–2835. https://doi.org/10.1002/2015GL063498.
St. Jacques, J.-M., Sauchyn, D.J., 2009. Increasing winter baseflow and mean annual streamflow from possible permafrost thawing in the Northwest Territories, Canada. Geophys. Res. Lett. 36, L01401. https://doi.org/10.1029/2008GL035822.
Standring, W.J.F., Stepanets, O., Brown, J.E., Dowdall, M., Borisov, A., Nikitin, A., 2008. Radionuclide contamination of sediment deposits in the Ob and Yenisey estuaries and areas of the Kara Sea. J. Environ. Radioact. 99, 665–679. https://doi.org/10.1016/j.jenvrad.2007.09.005.
Stephenson, S.R., Smith, L.C., Agnew, J.A., 2011. Divergent long-term trajectories of human access to the Arctic. Nat. Clim. Chang. 1, 156–160. https://doi.org/10.1038/nclimate1120.
Sutfin, N.A., Wohl, E., Dwire, K., 2016. Banking carbon: a review of organic carbon reservoirs in river systems. Earth Surf. Process. Landf. 41, 38–60. https://doi.org/10.1002/esp.3857.
Syvitski, J.P.M., 2002. Sediment discharge variability in Arctic rivers: implications for a warmer future. Polar Res. 21, 323–330. https://doi.org/10.3402/polar.v21i2.6494.
Tape, K., Sturm, M., Racine, C., 2006. The evidence for shrub expansion in Northern Alaska and the Pan-Arctic. Glob. Change Biol. 12, 686–702. https://doi.org/10.1111/j.1365-2486.2006.01128.x.
Toniolo, H., Kodial, P., Hinzman, L.D., Yoshikawa, K., 2009. Spatio-temporal evolution of a thermokarst in interior Alaska. Cold Reg. Sci. Technol. 56, 39–49. https://doi.org/10.1016/j.coldregions.2008.09.007.
Triska, F.J., 1984. Role of wood debris in modifying channel geomorphology and riparian areas of a large lowland river under pristine conditions: a historical case study. Verh Intern. Ver. Limnol. 22, 1876–1892.
Trumbore, S.E., Czimczik, C.I., 2008. An uncertain future for soil carbon. Science 321, 1455–1456. https://doi.org/10.1126/science.1160232.
Turcotte, B., Morse, B., Bergeron, N.E., Roy, A.G., 2011. Sediment transport in ice-affected rivers. J. Hydrol. 409, 561–577. https://doi.org/10.1016/j.jhydrol.2011.08.009.
U.S. Environmental Protection Agency, 2016. Climate Change Indicators in the United States, 2016, Fourth edition. No. EPA 430-R-16-004.
Urban, M., Forkel, M., Eberle, J., Hüttich, C., Schmullius, C., Herold, M., 2014. Pan-arctic climate and land cover trends derived from multi-variate and multi-scale analyses (1981–2012). Remote Sens. 6, 2296–2316. https://doi.org/10.3390/rs6032296.
Uunila, L.S., 1997. Effects of River Ice on Bank Morphology and Riparian Vegetation along Peace River, Clayhurst to Fort Vermilion.
Uunila, L., Church, M., 2015. Ice on peace river: effects on bank morphology and Riparian vegetation. In: The Regulation of Peace River. John Wiley & Sons, Ltd, Chichester, UK, pp. 115–140. https://doi.org/10.1002/9781118906170.ch6.
Vandenberghe, J., 2001. A typology of Pleistocene cold-based rivers. Quat. Int. 79, 111–121. https://doi.org/10.1016/S1040-6182(00)00127-0.
Vandenberghe, J., 2002. The relation between climate and river processes, landforms and deposits during the Quaternary. In: Quat. Int., Late Pleistocene and Holocene Investigations in Europe. International Conference on Past Global Changes. vol. 91. pp. 17–23. https://doi.org/10.1016/S1040-6182(01)00098-2.
Vandenberghe, J., 2008. The fluvial cycle at cold–warm–cold transitions in lowland regions: a refinement of theory. Geomorphology 98, 275–284. https://doi.org/10.1016/j.geomorph.2006.12.030.
Vandenberghe, J., Woo, M., 2002. Modern and ancient periglacial river types. Prog. Phys. Geogr. 26, 479–506. https://doi.org/10.1191/0309133302pp349ra.
Vonk, J.E., Gustafsson, Ö., 2013. Permafrost-carbon complexities. Nat. Geosci. 6, 675–676. https://doi.org/10.1038/ngeo1937.
Walvoord, M.A., Kurylyk, B.L., 2016. Hydrologic impacts of thawing permafrost—a review. Vadose Zone J. 15. https://doi.org/10.2136/vzj2016.01.0010.
Walvoord, M.A., Striegl, R.G., 2007. Increased groundwater to stream discharge from permafrost thawing in the Yukon River basin: potential impacts on lateral export of carbon and nitrogen. Geophys. Res. Lett. 34, 1–6. https://doi.org/10.1029/2007GL030216.
Walvoord, M.A., Voss, C.I., Wellman, T.P., 2012. Influence of permafrost distribution on groundwater flow in the context of climate-driven permafrost thaw: example from Yukon Flats Basin, Alaska, United States. Water Resour. Res. 48https://doi.org/10.1029/2011WR011595. n/a–n/a.
Wang, G., Hu, H., Li, T., 2009. The influence of freeze–thaw cycles of active soil layer on surface runoff in a permafrost watershed. J. Hydrol. 375, 438–449. https://doi.org/10.1016/j.jhydrol.2009.06.046.
Ward, J.V., Tockner, K., Arscott, D.B., Claret, C., 2002. Riverine landscape diversity. Freshw. Biol. 47, 517–539. https://doi.org/10.1046/j.1365-2427.2002.00893.x.
Wohl, E., 2013. Floodplains and wood. Earth-Sci. Rev. 123, 194–212. https://doi.org/10.1016/j.earscirev.2013.04.009.
Wohl, E., Dwire, K., Sutfin, N., Polvi, L., Bazan, R., 2012. Mechanisms of carbon storage in mountainous headwater rivers. Nat. Commun. 3 (1263). https://doi.org/10.1038/ncomms2274.
Wolman, M.G., 1959. Factors influencing erosion of a cohesive river bank. Am. J. Sci. 257, 204–216. https://doi.org/10.2475/ajs.257.3.204.
Wolman, M.G., Leopold, L.B., 1957. River Flood Plains: Some Observations on their Formation (no. 282–C). U.S. Geological Survey.
Woo, M., 1986. Permafrost hydrology in North America. Atmosphere-Ocean 24, 201–234.
Woo, M.-K., Thorne, R., 2003. Streamflow in the Mackenzie Basin, Canada. Arctic 56, 328–340.
Woo, M.-K., Winter, T.C., 1993. The role of permafrost and seasonal frost in the hydrology of northern wetlands in North America. J. Hydrol. 141, 5–31. https://doi.org/10.1016/0022-1694(93)90043-9. Introduction to the 28th International Geological Congress Symposium on the Hydrogeology of Wetlands.
Woo, M.-K., Kane, D.L., Carey, S.K., Yang, D., 2008a. Progress in permafrost hydrology in the new millennium. Permafr. Periglac. Process. 19, 237–254. https://doi.org/10.1002/ppp.613.
Woo, M.-K., Thorne, R., Szeto, K., Yang, D., 2008b. Streamflow hydrology in the boreal region under the influences of climate and human interference. Philos. Trans. R. Soc. Lond. Ser. B Biol. Sci. 363, 2249–2258. https://doi.org/10.1098/rstb.2007.2197.
Wright, N., Hayashi, M., Quinton, W.L., 2009. Spatial and temporal variations in active layer thawing and their implication on runoff generation in peat-covered permafrost terrain. Water Resour. Res. 45. https://doi.org/10.1029/2008WR006880.
Wu, P., Wood, R., Stott, P., 2005. Human influence on increasing Arctic river discharges. Geophys. Res. Lett. 32, L02703. https://doi.org/10.1029/2004GL021570.
Wynn, T.M., Henderson, M.B., Vaughan, D.H., 2008. Changes in streambank erodibility and critical shear stress due to subaerial processes along a headwater stream, southwestern Virginia, USA. Geomorphology 97, 260–273. https://doi.org/10.1016/j.geomorph.2007.08.010.
Yamazaki, Y., Kubota, J., Ohata, T., Vuglinsky, V., Mizuyama, T., 2005. Seasonal changes in runoff characteristics on a permafrost watershed in the southern mountainous region of eastern Siberia. Hydrol. Process. 20, 453–467. https://doi.org/10.1002/hyp.5914.
Yoshikawa, K., Bolton, W.R., Romanovsky, V.E., Fukuda, M., Hinzman, L., 2002. Impacts of wildfire on the permafrost in the boreal forests of Interior Alaska. J. Geophys. Res. Atmos. 107, FFR 4-1. https://doi.org/10.1029/2001JD000438.
Young, K.L., Woo, M., Edlund, S.A., 1997. Influence of local topography, soils, and vegetation on microclimate and hydrology at a high Arctic Site, Ellesmere Island, Canada. Arct. Alp. Res. 29, 270. https://doi.org/10.2307/1552141.
Yumoto, M., Ogata, T., Matsuoka, N., Eiji, 2006. Riverbank freeze-thaw erosion along a small mountain stream, Nikko volcanic area, Central Japan. Permafr. Periglac. Process. 17, 325–339. https://doi.org/10.1002/ppp.569.
Zhang, T., Barry, R.G., Knowles, K., Ling, F., Armstrong, R.L., 2003. Distribution of seasonally and perennially frozen ground in the Northern Hemisphere. In: Phillips, M., Springman, S.M., Arenson, L.U. (Eds.), Permafrost: Proceedings of the 8th International Conference on Permafrost, Zurich, Switzerland, 21–25 July 2003. A.A. Balkema Publishers, Brookfield, Vt, pp. 1289–1294.
Zhang, T., Frauenfeld, O.W., Serreze, M.C., Etringer, A., Oelke, C., McCreight, J., Barry, R.G., Gilichinsky, D., Yang, D., Ye, H., Ling, F., Chudinova, S., 2005. Spatial and temporal variability in active layer thickness over the Russian Arctic drainage basin. J. Geophys. Res. Atmos. 110, D16101. https://doi.org/10.1029/2004JD005642.
Zhang, W., Miller, P.A., Smith, B., Wania, R., Koenigk, T., Döscher, R., 2013. Tundra shrubification and tree-line advance amplify arctic climate warming: results from an individual-based dynamic vegetation model. Environ. Res. Lett. 8, 034023. https://doi.org/10.1088/1748-9326/8/3/034023.
Zimov, S.A., Davydov, S.P., Zimova, G.M., Davydova, A.I., Schuur, E.a.G., Dutta, K., Chapin, F.S., 2006. Permafrost carbon: stock and decomposability of a globally significant carbon pool. Geophys. Res. Lett. 33. https://doi.org/10.1029/2006GL027484.