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Authors: Crawford, John T., and Stanley, Emily H.

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Distinct Fluvial Patterns of a Headwater Stream Network Underlain by Discontinuous Permafrost

John T. Crawford†‡ and Emily H. Stanley**

*Center for Limnology, University of Wisconsin–Madison, 680 North Park Street, Madison, Wisconsin 53706-1413, U.S.A.
†National Research Program, U.S. Geological Survey, Boulder, Colorado 80303, U.S.A.
‡Corresponding author: jtcrawford@wisc.edu

Abstract

Hydrology, permafrost, and vegetation will likely respond to warming of northern latitudes with concurrent shifts in channel form and network pattern. At present, data on channel structure and networks in most northern regions are sparse, thus restricting any comprehensive understanding of channel processes or predictions of change in response to warming. We conducted a survey in 2011 of stream hydraulic geometry and network pattern in an upland headwater catchment of the Yukon River basin that is underlain by discontinuous permafrost. We found atypical hydraulic geometry exponents for depth and velocity but not for width and slope. We also found the study catchment to have unusually low drainage density and bifurcation ratios. Our data support the hypothesis that snow and channel ice decrease geomorphic effectiveness during snowmelt, which ultimately constrains channel and network development. Additionally, qualitative data support the hypothesis that dense riparian vegetation promotes bank stabilization and leads to nearly vertical channel walls in small streams, thus leading to anomalous hydraulic geometry. Simple metrics such as the drainage density and hydraulic geometry relationships may prove to be useful metrics of boreal and subarctic warming and permafrost thaw, and this study may serve as an important baseline to evaluate future changes.

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Introduction

Arctic and subarctic channel networks often have unusual physical features and can depart from conventional alluvial forms (McNamara et al., 1999; Boucher et al., 2009). Permafrost, peat soils, snow, ice, and freeze-thaw cycles have all been identified as drivers that often create distinct channel and drainage forms because of their ability to enhance or limit sediment mobilization, transport, and deposition (e.g., Watters and Stanley, 2007; Boucher et al., 2009; McNamara and Kane, 2009). However, the causes of, and the degree to which subarctic and arctic channel forms differ from typical conditions is difficult to determine, as data on fluvial structure and dynamics are sparse for high latitude boreal and arctic regions (Priesnitz and Schunke, 2002; Järvelä and Helmiö, 2004; Best et al., 2005; McNamara and Kane, 2009; Baki et al., 2012).

Warming air temperatures and thawing of permafrost could significantly alter stream network patterns and morphology across a vast portion of the globe. Permafrost is a defining physical feature of northern boreal, subarctic, and arctic regions and covers roughly 24% of the northern hemisphere (Zhang et al., 2003). Permafrost and seasonally frozen ground create unique landscape forms such as patterned ground, pingos, soil waves due to gelifluction, and self-sorted stone stripes (Black, 1976). Permafrost regions are also associated with unusual drainage features such as water tracks and beaded drainages (Brown and Kupsch, 1974; McNamara et al., 1999), and unusual cross-sectional features due to channel ice (McNamara and Kane, 2009). These landscape forms may be sensitive to slower, sustained alterations (press perturbations) such as permafrost degradation or hydrologic change, in addition to instantaneous alterations such as fire or severe storms (pulse perturbations). Progressive cryosphere shrinkage in high latitude environments due to rapid climate warming is receiving substantial attention (Jorgenson et al., 2001; Hinzman et al., 2005), and several consequences of warmer conditions are beginning to emerge, including shortened winter ice coverage in lakes (Magnuson et al., 2000) and large-scale permafrost thaw in the Arctic Coastal Plain leading to increased coverage and density of thermokarst features (Jorgenson et al., 2006). But less is known about headwater channel networks. Thermokarst features can alter downstream sediment delivery, nutrient export (Bowden et al., 2008), and aspects of the carbon cycle in headwater basins via delivery and emissions of carbon gases (Crawford et al., 2013). At the larger scale in Alaska's Yukon River basin, permafrost thaw has altered water flowpaths (Walvoord and Striegl, 2007), with impacts on carbon cycling (Striegl et al., 2005), and has likely led to increased mercury export associated with particulates (Schuster et al., 2011). Changes in channel patterns and morphology could have consequences for elemental cycles and the ecology of stream ecosystems in many permafrost regions.

Given often-unusual channel forms relative to better-studied temperate regions, and the potential for large-scale ecological changes in response to permafrost thaw, the goal of this paper is to better understand fluvial morphological conditions in a region of discontinuous permafrost and to evaluate potential drivers of atypical network patterns. We asked three questions regarding the geomorphology of permafrost streams: (1) How does stream channel cross-sectional form adjust to varying water discharge in permafrost landscapes? (2) How effectively do streams drain and dissect a permafrost landscape? (3) How do these patterns relate to other regions of permafrost and dissimilar regions of the globe, and what might explain the divergence? While analyses of channel cross-section relationships and network structure are relatively simple, these techniques have proven useful when addressing novel fluvial forms such as bedrock channels (Montgomery and Gran, 2001), urban networks (Doll et al., 2002), and even extraterrestrial landscapes (Som et al., 2009). Thus, because of the paucity of studies in permafrost regions, we provide a first basic quantitative description of headwater stream channel form in Alaska's Yukon River basin by leveraging the hydraulic geometry approach along with other straightforward geomorphic techniques.

PRINCIPLES OF HYDRAULIC GEOMETRY AND NETWORK ANALYSIS

Stream channels are vectors of water and sediment transport from the continents to the ocean. The cross-sectional profile of a stream reach reflects the average magnitude of water and sediment discharge from relatively infrequent "channel-forming" conditions, which can be translated roughly to bank-full flow. Additional constraints imposed by the stream boundary, via cohesion of sediments and vegetation, provide varying levels of protection from erosion. The hydraulic geometry concept (Leopold and Maddock, 1953) is an empirical, statistical framework that can describe average adjustments of channel cross sections in response to variations in water discharge, which is thought to be the primary independent determinant of channel shape (Knighton, 1998). Hydraulic geometry relates planes of the channel cross section (wetted width [w, m], depth [d, m], and water velocity [v, m s^{-1}]) to discharge (Q; m^3 s^{-1}) through power law functions (Equations 1–3):

$$w = aQ^b, \tag{1}$$

$$d = cQ^f, \tag{2}$$

$$v = kQ^m. \tag{3}$$

Analysis of hydraulic geometry can be accomplished by monitoring individual stream cross sections over a range of discharge (the at-a-station approach), or by assessing the evolution of channel shape along a gradient of channel size (the downstream approach). General comparisons of hydraulic geometry focus on the exponents b, f, and m, and a large body of literature suggests that they conform to narrow ranges (b = 0.45–0.55; f = 0.2–0.4; m = 0.10–0.17; Knighton, 1998, and references therein; ASCE Task Committee, 1998; Millar, 2005). The three hydraulic geometry equations (1–3) are related to discharge through the continuity equation (Equation 4),

$$Q = w\, d\, v, \tag{4}$$

which requires that Equation 5 is also satisfied,

$$b + f + m = 1. \tag{5}$$

The concept of hydraulic geometry can be extended to other independent variables such as slope (s; Equation 6), resistance to flow (channel roughness and the Darcy-Weisbach friction factor), and sediment load by using similar power law equations and data on channel and basin properties:

$$s = gQ^z. \tag{6}$$

Mathematical representations of network patterns including the drainage density (D_d) and bifurcation ratio (R_B, also known as the Law of Stream Numbers) can be used to assess the degree to which stream channels dissect a landscape. Drainage density (km km^{-2}; Equation 7) varies widely across the globe in response to precipitation and landscape erosivity (Knighton, 1998). Notably, the response to precipitation is nonlinear. Values can range from ~1 to 20, with the highest variability found in basins receiving moderate annual precipitation (~400 mm yr^{-1}) (Gregory, 1976).

$$D_d = \Sigma L\, A_d^{-1} \tag{7}$$

where ΣL is the length of all channels in a drainage basin (km), and A_d is the total area of the drainage basin (km^2).

Similarly, the bifurcation ratio (unitless, Equation 8) describes the number of channels of order u relative to the next higher order (according to the Strahler classification system) in a basin of interest:

$$N_{u\text{-}1}\, N_u^{-1} \approx R_B \tag{8}$$

where N_u is the number of streams of order u. Unlike drainage density, the bifurcation ratio tends to converge on a narrow range of values (3–5; Smart, 1972).

REGIONAL SETTING

We conducted this study in the Nome Creek Basin, an upland headwater catchment in interior Alaska that is a tributary to Beaver Creek and the Yukon River (Fig. 1). The focus of this study was on small streams (1st order), but we include data that span the full range of stream size in the Nome Creek catchment (see Table 1). Mean annual temperature was –2.6 °C, and mean cumulative precipitation was 547 mm yr^{-1} for the period 2007–2011 measured at the Upper Nome Creek SNOTEL station run by the National Resources Conservation Service (65°22′N, 146°36′W, 768 m a.s.l.; adjacent to the Summer Creek site). Approximately one-third of annual precipitation in the basin occurs as snowfall. Mean elevation in the catchment is 623 m a.s.l., and maximum elevation is 1591 m a.s.l. Total drainage area of the Nome Creek outlet is approximately 443 km^2. Portions of the study region are above treeline and can be characterized as alpine tundra, with thin soils and areas of exposed regolith. Lower elevations, below treeline, tend to have significantly thicker soils (mostly gelisols). For a further description of soils in the region see Ping et al. (2005). Mosses (*Sphagnum* spp. and *Pleurozium* spp.) are the dominant ground cover and a layer of accumulated *Sphagnum* material is present (<30 cm depth). Thus, given one common definition (depth >30–40 cm; Gorham, 1991), true peat deposits are rare in the basin. Forested areas are dominated by *Picea mariana* (black spruce), although there are some patches of *Picea glauca* (white spruce) along the main stem of Nome Creek as well as isolated patches elsewhere in the catchment. Valley bottoms and riparian areas are dominated by *Salix* spp. (willow), *Ledum groenlandicum* (Labrador tea), *Vaccinium uliginosum* (blueberry), and other typical taiga flora. Particle sizes in channels are highly variable. First order streams have a mixture of small, flat cobbles and gravels and silt material, whereas higher order streams are dominated by smaller grain particles in pools and runs, but larger gravels and cobbles in riffle reaches. Riparian and bank sediments are typically silt loess. Precise bed and riparian particle sizes were not measured in this study.

The regional geology is predominantly Paleozoic metamorphic schists (Brabets et al., 2000). The area is part of the larger Yukon-Tanana terrane and was unglaciated throughout the Pleis-

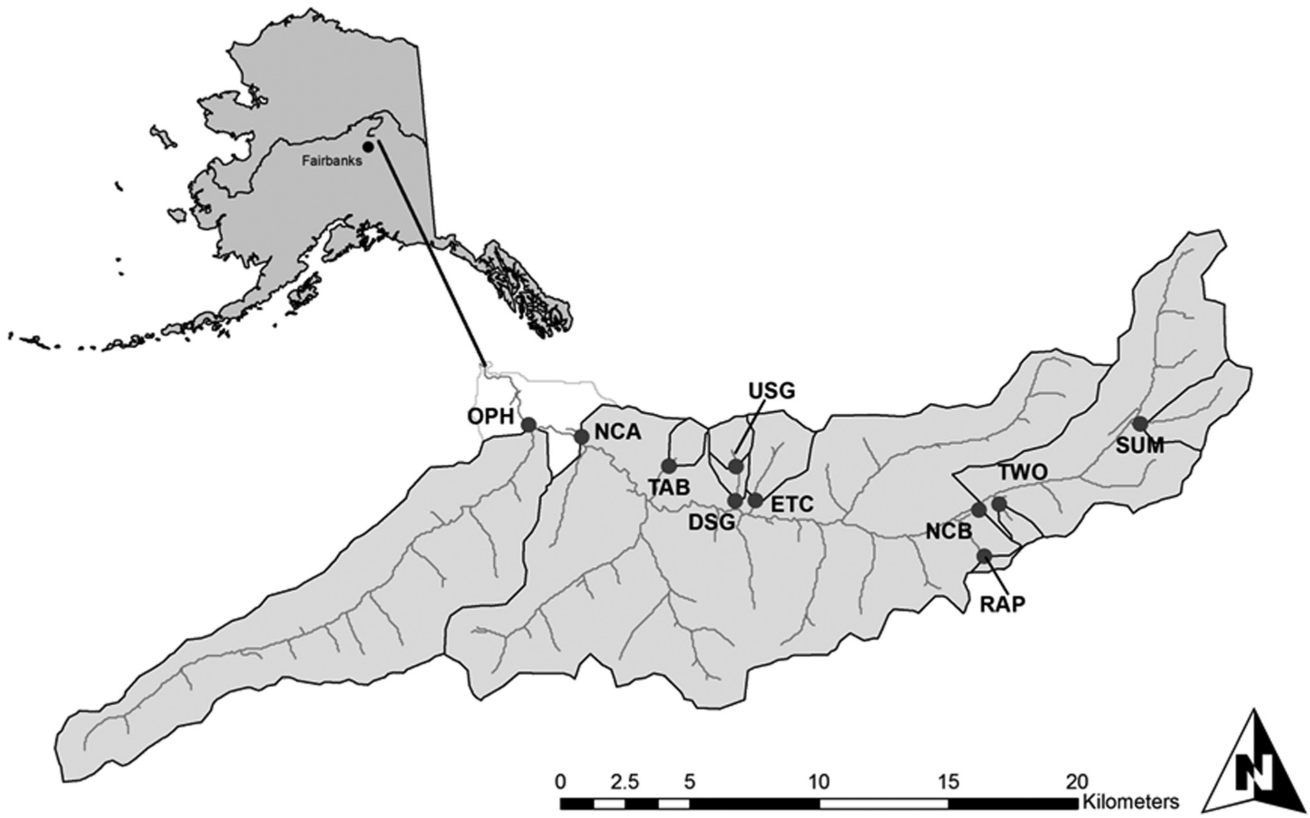

FIGURE 1. The Nome Creek study area in interior Alaska (65.34045°N, 146.713°W), indicating sites surveyed for channel cross section. See Table 1 for additional site details.

tocene (Petrone et al., 2006), which led to the formation of loess deposits throughout the region and rounded ridges that typically form V-shaped canyons with flat alluvial valleys (Brabets et al., 2000). It is also a zone of discontinuous permafrost which, in some locations, has partially thawed due to recent wildfire. However, overall extent of permafrost appears high, as even south-facing slopes that receive higher amounts of solar radiation are underlain by significant permafrost (based on soil pits, active layer probing,

TABLE 1

Characteristics of sampling locations from this study (site abbreviations correspond to locations shown in Fig. 1); USG and DSG sites are both located within the West Twin Creek catchment.

Stream name	Abbreviation	Contributing area (km^2)	Drainage density ($km\ km^{-2}$)	Channel slope (%)*	Basin slope (%)	Strahler stream order	Mean discharge ($m^3\ s^{-1}$)
Rapture Creek	RAP	0.74	0.27	6.53	3.7	1	0.005
Two Step Creek	TWO	1.46	0.79	8.82	7.5	1	0.02
Table Creek	TAB	2.15	0.159	10.11	5.3	1	0.01
West Twin Creek (upstream gage)	USG	2.6	0.354	7.68	5.7	1	0.059
West Twin Creek (downstream gage)	DSG	4.08	0.554	1.57	5.7	1	0.098
East Twin Creek	ETC	6.37	0.601	1.56	5.2	1	0.17
Summer Creek	SUM	6.87	0.574	6.78	6.3	1	0.278
Ophir Creek	OPH	81.38	0.58	1.04	0.7	2	1.025
Nome Creek (bridge crossing)	NCB	193.18	0.211	1.5	1.2	3	1.321
Nome Creek (administrative site)	NCA	362.09	0.416	0	1.2	4	n.d.

* 500 m upstream slope, except NCA, which is the mean slope of Nome Creek.

and also inferred from continuous coverage of black spruce and not *Betula neoalaskana* [paper birch]). Road construction in 2011 exposed some larger ice masses near the south-facing West Twin Creek (WTC), confirming the presence of significant ice-rich permafrost on south-facing slopes. Maximum active layer thickness on hillslopes is ~0.5 m (Kelsey et al., 2012; also see Petrone et al., 2007). The shallow active layer likely constrains flow through organic soils, especially when the active layer is not fully developed. Additionally, self-organized stone stripes on hillslopes may constrain water to flowpaths orthogonal to main channels. These areas are clearly visible from aerial imagery by having significantly greener vegetation. There is evidence of thermokarst formation in many locations within the study region, characterized by dead, tilting black spruce trees (commonly referred to as "drunken trees") and standing surface water in riparian areas and on burned hillslopes.

Materials and Methods

We used a variety of simple metrics to assess channel cross-section relationships and the drainage pattern of streams in the Nome Creek basin. These metrics were then compared to other permafrost networks and typical values for the globe. Basin slope, drainage area, drainage density (Equation 7), bifurcation ratios (Equation 8), and Strahler channel order were determined using Geographic Information Systems. Watersheds were delineated using stream lines from the National Hydrography Dataset (NHD) and a 60 m digital elevation model downloaded from http://ned.usgs.gov. We used multiple thresholds of minimum drainage area in the GIS delineation routines, until the locations and lengths of streams were consistent with our extensive field survey of stream locations and lengths. While the low resolution of the digital elevation model could lead to biased estimates of drainage density and other network metrics, observations from our field campaign generally confirmed the patterns obtained from GIS analysis. However, these results may still need to be interpreted with caution. Slope was calculated from the elevation change over a 500 m reach length above each survey location. Regular cross-section measurements were made to assess hydraulic geometry relationships of streams in the basin. Wetted widths, water depths, and water velocities were surveyed at four sites every 1–2 weeks, and two to three times at five additional sites between May and August 2011. Velocity was measured using a U.S. Geological Survey (USGS) Model 6205 pygmy current meter. The first four discharge measurements along West Twin Creek (DSG) were during snowmelt as water flowed over solid ice. Continuous discharge was monitored using stream pressure transducers and established rating curves at two locations along West Twin Creek beginning when the stream was fully contained within the channel (USG and DSG; described further in Crawford et al., 2013). Determination of bank-full flow was difficult at some locations, so cross-sectional width, depth, and velocity were measured based on the wetted channel form. Hydraulic geometry relationships (Leopold and Maddock, 1953) for width, depth, velocity, and slope (Equations 1–3 and 6) were determined by least square regression of log-transformed data. Because long-term gaging records do not exist at this remote location, and because the appropriate channel-forming flow is difficult to know at sites characterized by anchor ice (Priesnitz and Schunke, 2002; McNamara and Kane, 2009), we examined exponents extracted from the regression that used flow data from the day with highest recorded discharge at the site (Q_{max}), as well as from a second regression using data collected at all sites during summer baseflow (Q_{base}), the period between 26 July and 4 August 2011. In addition to basin-wide sampling, we measured discharge and channel cross sections along a longitudinal transect of West Twin Creek to look at small-scale evolution of channel form. Ancillary precipitation and temperature data were retrieved from the Upper Nome Creek SNOTEL station.

Results

Seasonal snowmelt in the Nome Creek catchment began in early May 2011 and was short-lived, with ice and snow essentially absent by 24 May 2011 (see Fig. 2). Many small catchments experience groundwater flow throughout the winter (P. F. Schuster, personal communication), which resulted in anchored bed ice in the channel as well as aufeis (surface-layered ice from groundwater discharge) in riparian areas. We found that aufeis cover in the West Twin Creek catchment was spatially related to groundwater inflows that occurred during the summer months. During most of the snowmelt period, streams did not occupy a distinct channel but instead meandered throughout the riparian zone, eroding

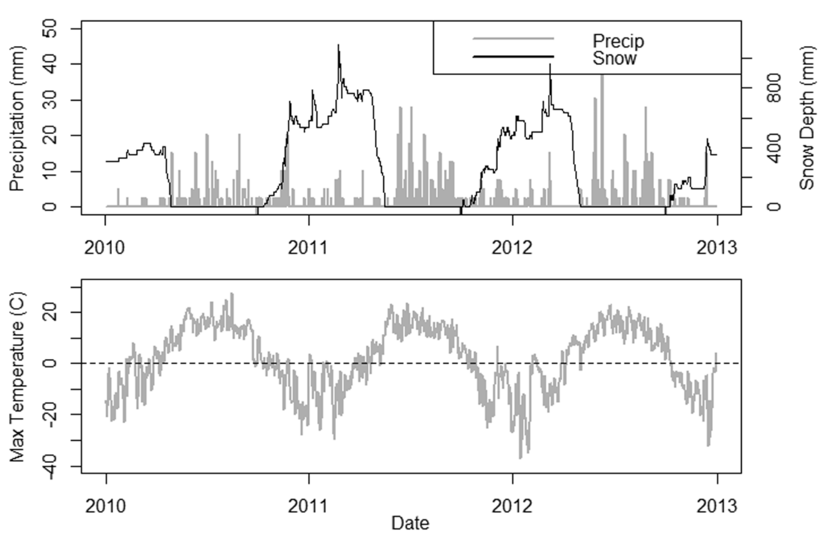

FIGURE 2. Precipitation, snow depth, and maximum air temperature time series from the Upper Nome Creek SNOTEL station.

FIGURE 3. West Twin Creek cutting down a temporary ice channel on 6 May 2011 during snowmelt. The overland location of the channel is apparent as water is flowing through a shrub meadow area.

aufeis and the remnant snowpack. This phenomenon of snowmelt runoff cutting through ice is shown at WTC in Figure 3. The temporary ice channel was ~50 m from the normal summertime channel. Complete down-cutting of anchored bed ice and reestablishment of defined channel flow occurred in less than 48 hours. This period of over-ice flow makes traditional monitoring or gaging of snowmelt runoff virtually impossible for many small streams in the catchment. Like many larger streams in the region, snowmelt along the main channel of Nome Creek was accompanied by raft ice of relatively small size.

Some of the highest measured discharges at WTC occurred over ice during snowmelt in May (Fig. 4), but we suspect that these early discharge values underestimate peak flows based on the timing and pace of snowmelt in 2011. Discharge data from USG and DSG sites on WTC revealed relatively low variability in flow once the stream had cut through anchor ice and returned to the channel (continuous record begins approximately 48 hours after channel was reestablished). Summer rainstorms were fairly regular (Fig. 2), but responses in discharge varied between the two sites on West Twin Creek, with little runoff occurring at USG (Fig. 4). The coefficient of variation (CV) at USG was 30%, and $Q_{\max}$ (0.126 m^3 s^{-1}) was just over two times greater than Q_{mean} (0.059 m^3 s^{-1}). Discharge magnitude and variability increased at DSG, but only slightly, reaching a CV of 57%, and Q_{max}/Q_{mean} of 3.4. The lower site experienced a maximum Q of 0.331 m^3 s^{-1}. In contrast, maximum measured flow during snowmelt at DSG over ice was 0.208 m^3 s^{-1}. The CV for flows at both sites fell well within average values for to the United States as a whole (Vogel et al., 2003).

LANDSCAPE PATTERN

We found unusual and distinctive landscape features in this permafrost stream network. The Nome Creek system was characterized by relatively long, narrow headwater catchments in which hillslopes were drained by a series of straight, parallel, often evenly spaced headwater channels that merged at approximate right angles into long, relatively low-gradient valley channels (see Fig. 1). Although the number of channels decreased with increasing stream order, bifurcation ratios were low for 1st/2nd and 3rd/4th order channels (2.1 and 1.4, respec-

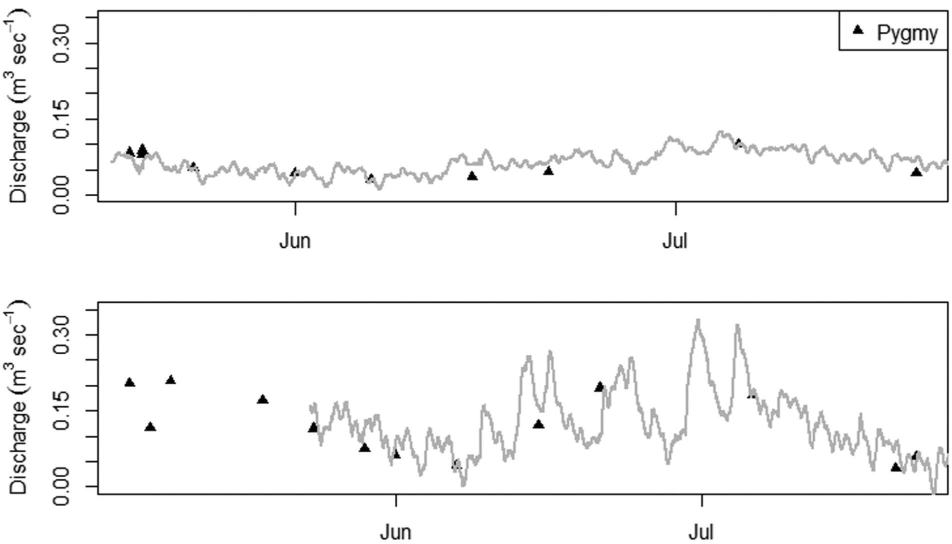

FIGURE 4. Discharge time series from West Twin Creek at the upstream gage (USG, top), and the downstream gage (DSG, bottom). The continuous record at both gages is for flow in the defined channel after snowmelt; pygmy measurements from 6 to 19 May 2011 at DSG were made over ice.

tively, compared to typical values of 3–5; Smart, 1972). Similarly, mean channel length did not increase with increasing stream order, reflecting the wide range of channel lengths within a given stream order (e.g., 1st order channels varied from <100 to >4500 m). Third and 4th order channels were characterized by pronounced meandering, distinct riffle-pool structure, and often well-developed point and in-channel bars composed of large-caliber sediments.

DRAINAGE DENSITY

A second unusual and distinctive basin-scale feature of the Nome Creek system was its extremely low drainage density. Drainage density increased as a function of basin slope for the six 1st order catchments (r^2 = 0.57, p < 0.08; the USG site in West Twin Creek was excluded so that this drainage was not represented twice in the regression), but nonetheless never exceeded 0.8 km km^{-2}. Given the limited resolution of the DEM used in our GIS analysis, the true value may be higher, but our extensive field observations tended to support the finding of low drainage density. This low value contrasts with minimum values of ~5 km km^{-2} for regions with similar precipitation (Gregory, 1976). The entire 4th order Nome Creek catchment was drained by just over 400 m of channel per km^2 (Table 1). At a larger spatial scale, a regional map illustrates that these Nome Creek patterns of parallel hillslope drainage, right-angle confluences with valley channels, and low drainage density were present in other surrounding catchments (Fig. 5). In contrast to some arctic landscapes, 0th order streams (also known as water tracks) were essentially absent. The only notable exception was an ephemeral, flowing wetland within a hillslope depression (between two large stone stripes) in the West Twin catchment. This and other depressions, which may contain surface water episodically, are distinctly visible in aerial imagery and could be misinterpreted as 0th order streams.

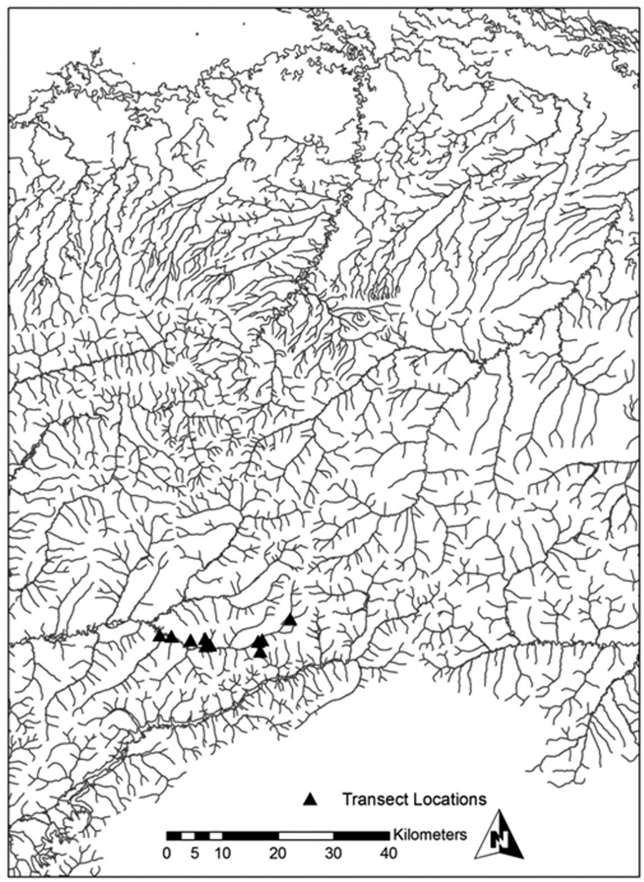

FIGURE 5. Drainage patterns in the Nome Creek region from the U.S. Geological Survey National Hydrography Dataset. Transect locations correspond to sites in Figure 1 and Table 1.

HYDRAULIC GEOMETRY

The hydraulic geometry exponent describing channel width (b, 0.5) among sites (downstream approach) corresponds well with established values for alluvial channels (0.45–0.55; ASCE Task Committee, 1998; Millar, 2005). Exponents were similar for equations based on Q_{max} and Q_{base} (Table 2), and the slopes and intercepts were not significantly different for regressions using these two different discharge values (test of homogeneity

TABLE 2

Fitted downstream hydraulic geometry exponents for the Nome Creek Basin (n = 10 sites).

Parameter	Exponent from Q_{max}	r^2	Exponent from Q_{base}	r^2
Width (*b*)	0.5	0.89	0.5	0.78
Depth (*f*)	0.19	0.62	0.28	0.69
Velocity (*m*)	0.3	0.82	0.23	0.59
Slope (*z*)	–0.34	0.52	–0.37	0.59

of slopes and ANCOVA, $p > 0.10$ in all cases). The exponent for slope (*z*; –0.34) also corresponded well with common alluvial channel values (generally –0.2 to –0.4; Knighton, 1998, and references therein). In contrast to width and slope, depth and velocity exponents in the Nome Creek (NC) drainage departed from expected equilibrium conditions. The depth exponent was lower (*f*, 0.19) than the typical range of 0.30–0.45. Similarly, the velocity exponent (*m*, 0.23–0.3) was higher than typical values for nonpermafrost regions (0.10–0.17; Knighton, 1998). The Nome Creek exponents were very similar to those reported by McNamara and Kane (2009) for the upper Kuparuk River. Cross-sectional profiles from the repeated gaging sites and one additional site (OPH), to represent a 2nd-order channel, indicate the narrow, deep nature of 1st order channels (RAP, USG, DSG) that respond to discharge increases by pronounced vertical adjustment, and then by increasing channel width in higher order channels (OPH and NCB). Reflective of this difference between narrow and deep headwater channels and widening at higher order sites, width:depth ratios across the study basin ranged from 1.1 to 21.7, but along a 1 km longitudinal transect in the 1st order West Twin Creek drainage beginning at the USG site, this ratio only varied from 1.1 to 6.0 and lacked any clear downstream trend (data not shown).

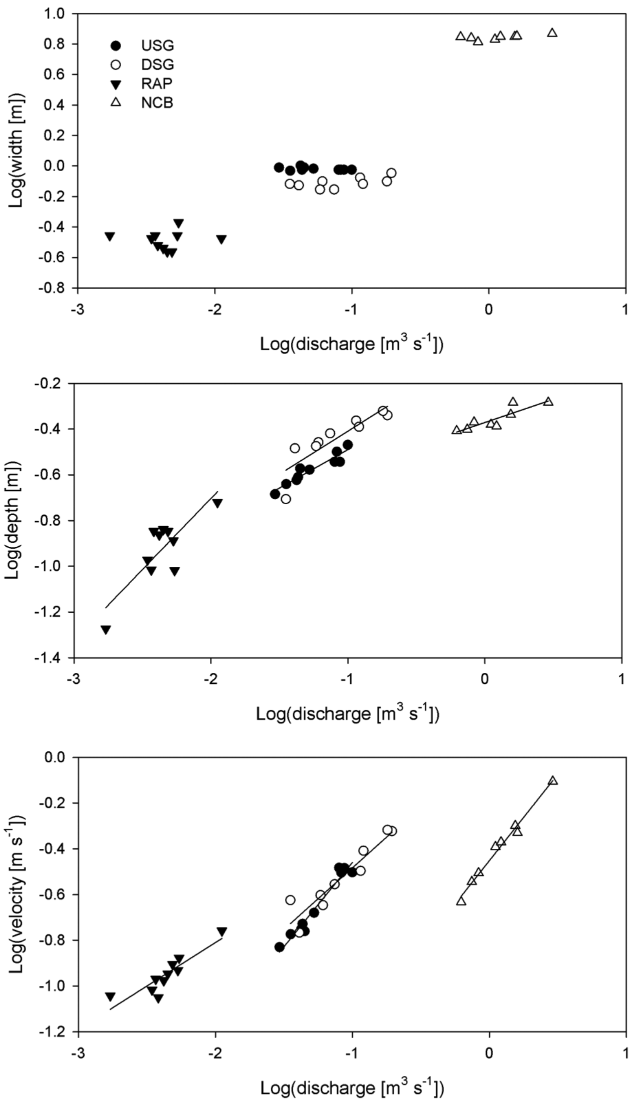

FIGURE 6. At-a-site relationships between discharge and channel width (top), depth (middle), and velocity for upstream (USG) and downstream (DSG) sites along West Twin Creek (WTC), Rapture Creek (RAP), and Nome Creek (NCB). Regression lines are shown for significant relationships between discharge and the channel variable. Exponent values and regression coefficients are listed in Table 3.

At-a-site relationships provided additional insights into cross-sectional forms by examining changes in wetted channel form in response to flow variation during summer months. Width did not respond to changes in flow among the four repeated survey sites (Fig. 6, part a; Table 3; $p > 0.05$). Instead, increases in discharge resulted in increased water depth and velocity (Fig. 6, parts b and c; Table 3). The response was similar among all sites except between the smallest (RAP) and largest streams (NCB) (LS Means comparison, $p < 0.05$), demonstrated by greater change in depth with increasing discharge at RAP, and a faster change in velocity at NCB.

Discussion

The aquatic network of the Nome Creek basin displayed many departures from typical alluvial conditions at both basin and channel scales. Notable patterns at the basin scale included the pinnate or trellis-like drainage form with straight headwater channels that joined valley channels at right angles, and distinctly low drainage density. Parvis (1950) suggested that eolian processes often create parallel ridge and valley forms within loess areas, and this is consistent with the drainage patterns and loess deposits observed in the Nome Creek study area, rather than reflecting local or regional joints and faults as suggested by Twidale (2004). Within this study region, well-developed meandering in lower gradient valley areas was not surprising. However, much straighter 1st order channels and low drainage density of the Nome Creek region was unexpected. Hillslope channels were extremely rare in many regions (e.g., the northern side of the OPH catchment; Fig. 1). Similarly, low drainage densities have also been reported in areas of discontinuous permafrost in northern Finland (Luoto, 2007) and Canada

TABLE 3

Fitted at-a-station hydraulic geometry exponents for sites shown in Figure 6. Results are reported for significant relationships only. See Table 1 for site name abbreviations.

Site	Depth (*f*)	r^2	Velocity (*m*)	r^2
USG (n = 10)	0.34	0.9	0.74	0.96
DSG (n = 13)	0.38	0.75	0.54	0.88
RAP (n = 10)	0.62	0.69	0.38	0.78
NCB (n = 8)	0.2	0.78	0.76	0.98

(Quinton et al., 2003). In these latter examples, low drainage densities were associated with abundant peat wetlands and low relief, which in turn were associated with high infiltration and/or storage capacity. The Nome Creek landscape with its steeper hillslopes and lack of well-developed peat wetlands represents a distinct departure from these two cases.

DRAINAGE DENSITY RELATIONSHIPS

Low drainage density is associated with a number of basin features, including high infiltration capacity, extremely arid conditions in which the rarity of precipitation limits channel development, and, in general, circumstances in which the erosivity of runoff is far less than the resistivity created by vegetation and parent lithology (Collins and Bras, 2010). With respect to infiltration capacity, our a priori expectation was that the widespread occurrence of permafrost and presence of regolith, coupled with steep hillslopes, would foster rapid supra-permafrost flow and overland flow and thus high channel formation potential. This was confirmed on multiple occasions when we observed overland flow and artesian discharge from soil pipes in headwater catchments during summer rainstorms. Low active layer storage during wet conditions and overland flow plus preferential flow through soil pipes (Koch et al., 2013) do not suggest high infiltration capacity. Therefore, it is likely that summer rainstorms could create channel-forming conditions in receiving streams. However, total precipitation in this region is relatively modest (<500 mm y^{-1}, one-third as snow), and evapotranspiration (ET) demand generally is high enough between storms to significantly decrease stored precipitation, especially near the summer solstice when ET is predicted to be highest (Hinzman et al., 2006). Thus, we expect that channel-forming conditions are limited for much of the period of time when ice is also absent. Extensive coverage of *Sphagnum* spp. and other mosses may also serve to decrease the energy of runoff and potentially limit soil compaction (Gray and Lieser, 1982; Coppin and Richards, 1990), thus reducing erosivity. Despite steep slopes and soil/geological conditions that should favor surface runoff and drainage evolution, low precipitation and widespread moss cover appear to be sufficient to counteract these physical conditions and limit channel development.

ANOMALOUS HYDRAULIC GEOMETRY

Channel cross sections departed from typical alluvial form in this headwater catchment. Basinwide variation in width, depth, and velocity exponents (*b, f*, and *m*) indicated that channel adjustments to increasing flow were in line for width (*b*), but not depth (*f*). In turn, limited depth adjustments were compensated by greater velocities (*m*). We can compare these findings with hydraulic geometry relationships documented for other boreal, arctic, and Antarctic regions (Table 4). Generally, the Nome Creek exponents are most similar to those of the Kuparuk River in arctic Alaska (McNamara and Kane, 2009) and rivers of southern Finland (Helmiö, 2004). Based on a limited number of studies, it appears that anomalous hydraulic geometries of the Nome Creek basin are in line with other ice and permafrost impacted systems. We contend that the "anomalous" nature of permafrost drainage networks requires researchers to adopt a different morphological context at the beginning of any study in these regions.

Representative channel cross-sectional profiles (Fig. 7) suggest a two-stage process of channel adjustment that is not necessarily apparent from the hydraulic geometry analysis. Specifically, increases in flow in 1st order channels were associated with

TABLE 4

Hydraulic geometry exponents for some boreal, arctic, and Antarctic environments; the Upper Kuparuk River values are for catchments <300 km².

Location	Width (*b*)	Depth (*f*)	Velocity (*m*)	Reference
Kuparuk River	0.5	0.2	0.16	McNamara and Kane (2009)
Northwest Territories, Canada*	0.14 (0–0.40)	0.17 (0.10–0.27)	0.65 (0.45–0.85)	Baki et al. (2012)
Southern Finland (rivers)	0.09–0.34	0.07–0.45	0.31–0.72	Helmiö (2004)
Northern Wisconsin (stream/fen)	–0.79	0.4	1.5	Watters and Stanley (2007)
Deception Island, Antarctica	0.1	0.53	0.36	Inbar (1995)
Interior Alaska (streams)	0.5	0.19	0.3	This study

*Mean for 6–11 streams (range).

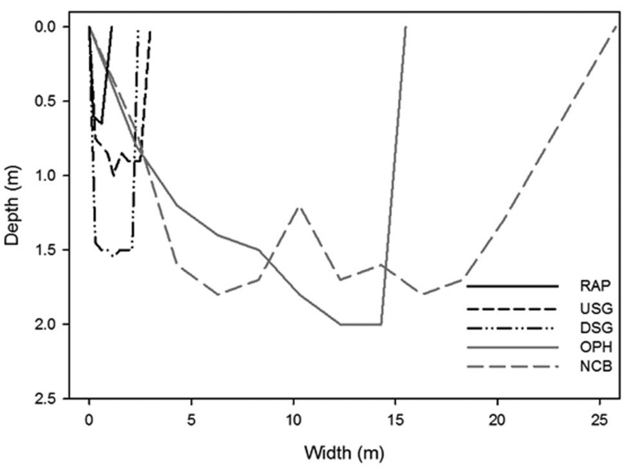

FIGURE 7. Channel cross sections for at-a-site monitoring stations of three 1st-order streams (USG, DSG, RAP), a 2nd-order stream (OPH), and the 3rd-order location of Nome Creek (NCA).

deepening but not widening, followed by widening accompanied by only slight deepening at more downstream, higher order sites. Klein (1981) described a similar shift in channel adjustment, suggesting that high flashiness (or "peakedness," quantified as the ratio of mean:maximum discharge) in small basins favors over-adjustment of channel depth, and subsequent decreases in flashiness favor greater width adjustment in larger basins. However, Nome Creek channels were not characterized by substantial flashiness (and in fact, for one drainage, flashiness increased with distance downstream), and all sites considered in this study except NCA fell within the "small" basin size. Thus, the flashiness of the flow regime and changes in basin size cannot explain this two-step adjustment pattern.

The vertical banks of 1st order sites and vertical, rather than lateral adjustment in Nome Creek channels could be a result of scour resistance associated with dense bank vegetation (predominantly willows) lining most channels. Dense rooting structures were visible along steep exposed banks in many of these headwater locations. The observed deep, narrow channel morphology with near-vertical banks has also been reported in a number of settings in association with well-developed root structure of plants such as grasses and dense riparian forest communities in temperate regions (reviewed by Anderson et al., 2004). Similarly, the potentially ice-rich loess sediments in riparian areas could have high cohesion that may protect channels from widening. While we do not have adequate data to test the hypothesis that sediment cohesion drives vertical channel banks, moderate cohesion has been documented for desiccated loess sediments in interior Alaska (Johnson and Lorenz, 2000). However, other geotechnical data suggest that ice-rich silts are actually quite erosive (Kanevskiy et al., 2012), and their hydraulic properties could contribute to channel incision in permafrost stream networks (Koch et al., 2013). The role of channel cohesion and resistance to scour from plants and sediment is suggested by the scatter in the hydraulic geometry relationships, as discharge alone cannot fully explain the observed forms (r^2 values between 0.52 and 0.89). These observations of potential bank stability are qualitative, and clearly more research into the geotechnical properties of these bank sediments and the role of vegetation is needed to understand the potential for future erosion in permafrost networks.

GEOMORPHIC EFFECTIVENESS

In addition to apparent bank stabilization by vegetation and perhaps strong sediment cohesion, departure from typical alluvial conditions was likely promoted by the pattern of snow and ice melt in the region. Ice is often a major geomorphic agent via scouring during the breakup period, causing over-widened channels (Boucher et al., 2009), but in these headwater basins, its presence might play a more protective role. Spring snowmelt and stream flow first begin with unconstrained water flow over anchor ice for many days before it cuts through this ice layer and returns to the regular, summertime channel, which we found could be up to 50 m away from the ice channel. The melt phase typically produces some of the highest annual flow rates, but high flows have no geomorphic effectiveness because the channel is shielded by ice. This phenomenon of high flows with limited capacity to do geomorphic work because of ice has also been described in headwater arctic basins of Alaska (McNamara and Kane, 2009) and northwestern Canada (Priesnitz and Schunke, 2002). In these cases, channel-forming discharges were associated with summer rainstorms that produced peak flows of lesser magnitude than during the snowmelt period. This may also be the case for Nome Creek, but we do not know the effectiveness of these larger events. Runoff responses to high intensity storms were observed to be variable across the Nome Creek basin. For example, in the WTC catchment, runoff responses following storms were often absent at the USG site but present at the lower elevation DSG site. Further, although "flashiness" increased downstream, variability in Q was small in the Nome Creek basin, highlighting a limited capacity for high summer flows to mobilize and transport sediments in a predictable fashion across all stream orders. Our data support the hypothesis of limited geomorphic effectiveness due to timing of flow events, but important geotechnical data and assessment of sediment transport in permafrost streams are greatly needed to fully understand the frequency and timing of erosion.

Conclusions

Given the low drainage density and atypical channel forms in this subarctic drainage, if channels become less protected from work by higher flows, then we might expect to see substantial channel adjustment and perhaps also drainage density expansion. Others have previously suggested that arctic networks will respond to atmospheric warming because sediment load is positively correlated with mean annual temperature (Syvitski, 2002). However, the direction of change for aspects such as the drainage density depends on the initial state of the system, and networks may actually contract during nonpermafrost conditions (and expand under permafrost) due to a decrease in sediment yield (Bogaart et al., 2003). Regardless of the direction, permafrost landscapes are changing; riparian thaw features observed in the Nome Creek system as well as in many other boreal and arctic sites (Lantz and Kokelj, 2008; Osterkamp, 2007) provide a clear geomorphic signal of this change. In addition to permafrost thaw, channel responses could be expected from factors such as decreases in runoff variability in headwater streams in the region (Jones and Rinehart, 2010) or decreased erosion due to riparian shrub expansion as has been witnessed in the Arctic (Tape et al., 2011). We suspect that potential changes in the timing and magnitude of snowmelt may also lead to changes in channel and network form due to the protective role of channel ice in these catchments. Alluvial changes are likely already progressing in permafrost regions and could increase with accelerating thaw. Therefore, this study should serve as an important baseline for evaluating the future effects of climate change in this and other subarctic catchments. Importantly, future fluvial geomorphic studies in permafrost regions require a conceptual framework that is significantly different from temperate alluvial systems. This is exemplified by unconventional hydraulic geometry relationships as well as low drainage density. Simple metrics such as the drainage density—which can be observed remotely—and hydraulic geometry relationships may prove to be useful metrics of future boreal and subarctic warming and permafrost thaw.

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