2018

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Geoscience Frontiers 9 (2018) 1193–1202

Contents lists available at ScienceDirect

China University of Geosciences (Beijing)

Geoscience Frontiers

journal homepage: www.elsevier.com/locate/gsf

Research Paper

Hillslope coupled stream morphology, flow conditions, and their effects on detrital sedimentology in Garnet Canyon, Teton Range, Wyoming

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A R T I C L E I N F O

Article history:
Received 14 April 2017
Received in revised form
28 June 2017
Accepted 18 July 2017
Available online 8 August 2017
Handling Editor: M. Santosh

Keywords:
Stream
Morphology
Garnet Canyon
Sediment
Talus

A B S T R A C T

Characterizing stream erosion in any steep mountain landscape is arduous, but the challenge level increases when the stream flows through a glaciated catchment frequently modified by hillslope debris. Glacial landforms and stochastic mass wasting in alpine systems may interfere with sediment delivery to downstream sites where detrital sediments are often collected to represent upstream bedrock sources. To use detrital sediments as indicators of erosion, we need to understand potential sediment accumulation in flat glaciated reaches or behind rockfall barriers. This study investigates the stream channel in Garnet Canyon, a glaciated catchment located in the central Teton Range, to describe hillslope coupled channel morphology and the subsequent effects on sediment transport throughout the catchment. Stream cross-section surveys and sediment size measurements of the surface bedload were collected in the field within a glacially flattened segment of Garnet Canyon. Calculations of shear stress conditions allowed evaluation of the importance of mineral densities on potential grain entrainment. The length of the Garnet Canyon stream observed in this study was coupled with hillslope deposits. Critical shear stresses were sufficient to move gravel-sized sediments through all sections when calculated with quartz mineral density and through most sections when applying apatite mineral density. These results verify the application of detrital sediments to evaluate erosion rates or spatial bedrock sources because snowmelt stream flow efficiently moves entrained sediment past glacially reduced slopes and potential talus barriers.

1. Introduction

Steep mountain streams play an important role in shaping landscapes and delivering water and sediments to larger, downstream rivers. The capacity for mountain streams to erode bedrock and transport sediments relies on channel morphology, underlying bedrock or sediments, and discharge (Whitbread et al., 2015). Channel morphology, from upstream to downstream, varies depending on glacial and hillslope conditions (Montgomery and Buffington, 1997; Sklar and Dietrich, 2001; Halwas and Church, 2002; Chin and Wohl, 2005; Brardinoni and Hassan, 2006, 2007; Norton et al., 2008). Step-pools, for example, reduce a stream's energy, thereby reducing the stream's ability to erode bedrock and transport sediments (Chin and Wohl, 2005). In post-glacial systems, however, channel morphology distributions are complicated by spatial variability of glacial incision. Glacial deepening creates localized areas in alpine settings with decreased slopes and increased potential for sediment storage (Alley et al., 2003; Dühnforth et al., 2008; Straumann and Korup, 2009). Additionally, glaciers destabilize hillslopes, promoting mass wasting that results in significant accumulation on the valley floor and potential barriers to stream flow (Tranel and Strow, 2017). Sediment accumulation from hillslope debris creates localized changes in slope and substrate, including the size of sediments deposited, that influence stream efficiency (Chin and Wohl, 2005; Norton et al., 2008). Studies that apply detrital mineral analyses to understand catchment averaged erosion rates, trace spatial patterns of erosion, or approximate age-elevation profiles and timing of orogenic events rely on efficient streams to transport a range of sediment grain sizes to their mouths (Stock et al., 2004, 2006; Avdeev et al., 2011; Tranel et al., 2011; Ehlers et al., 2015).

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Peer-review under responsibility of China University of Geosciences (Beijing).

http://dx.doi.org/10.1016/j.gsf.2017.07.005

A stream's ability to transport sediments throughout a catchment has important implications for detrital studies. If sediments are trapped behind a barrier upstream, then sediments collected at the mouth of the stream only represent a fraction of the entire catchment downstream from the barrier. Additionally, sediments trapped upstream prevent incision into the valley floor (Norton et al., 2008). Riebe et al. (2015) observed that sand and gravels indicate different source locations for sediments collected at the mouth of a stream. For example, at their field site they observed that sands derive from lower elevations than gravels. With evidence that different sized sediment fractions produce varying results in age distributions and catchment averaged erosion rates, it is further important to investigate upstream hydraulic variability to explain from where sediments are derived and how upstream conditions in various channel morphologies may influence the transport of sediments throughout a mountain system (Riebe et al., 2015; Foster and Anderson, 2016; Lukens et al., 2016). Additionally, it is important to consider the density of the detrital mineral systems, which affects how grains are entrained or deposited. Most studies investigating sediment transport in stream systems apply the density of quartz minerals. Quartz is abundantly used in cosmogenic nuclide analyses, however, apatite and zircon minerals are used to evaluate spatial patterns of erosion. The greater densities of apatite, zircon, or other heavy minerals composing gravel sediments may influence their transport trajectories through catchments (Garzanti et al., 2009).

The purpose of this study is to investigate the potential for sediment transport in typical summer snowmelt discharges through Garnet Canyon in the Teton Range, Wyoming and discuss how accurately sediments captured at the mouth of the canyon represent the entire upstream system. This research investigates two questions. First, how do glaciers and rockfalls influence stream reach morphology in a small to moderately sized, alpine catchment? Second, how efficiently are sediments transported throughout different classes of stream reaches influenced by glacial erosion and active hillslope processes? Garnet Canyon contains an alpine stream flowing across a glacially eroded basin displaying an abundance of talus debris along much of its length. This study evaluates channel observations and flow conditions to determine if sands and gravels are transported throughout the system or trapped due to low slope or barriers related to glaciation or hillslope failures. Bedload sediment sizes were compared to detect upstream to downstream progression in grain size distributions. By characterizing the stream morphology and hydraulic conditions in a single catchment that was previously studied with detrital sediments, it is possible to assess the need for similar and more intensive observations in other alpine canyons or regions for future sediment transport and detrital analyses.

2. Geological setting

The Teton Mountains are a relatively small mountain range located in western Wyoming, south of the Yellowstone volcanic high. Composed of Archean metamorphic basement and Paleozoic to Mesozoic sedimentary rock, these mountains have experienced a complex tectonic history related to Laramide thrusting, Basin and Range extension, and Yellowstone volcanism (Foster, 1947; Reed and Zartman, 1973; Craddock et al., 1988; Roberts and Burbank, 1993; Smith et al., 1993; Byrd et al., 1994; Zartman and Reed, 1998; Frost et al., 2006; Pickering-White et al., 2009). Pleistocene glaciers subsequently carved deep canyons into the bedrock and created steep hillslopes (Pierce and Good, 1992; Foster et al., 2010). Glaciers flowed through east draining canyons at least three times within the last 100 ky. During the last glacial maximum, the mountain glaciers extended into Jackson Hole at the foot of the mountains ∼11 ka (Licciardi and Pierce, 2008; Larsen et al., 2016).

The relatively low-lying Snake River Plain, located east of the Teton Range, directs moist air currents and precipitation (snowfall) into the mountains. Winds from the west and northwest carry precipitation over the peaks. Average snowfall measured near the Teton Mountains is 4–5 m per year (Dirks and Martner, 1982) and mean annual precipitation is 120–201 cm. Snow blowing over ridges accumulates in shaded rock crevices and cirques below east-facing peaks (Foster et al., 2010). Temperatures below high peaks remain cool at high elevations, allowing preservation of winter snowpack and small glaciers through the summer months (Love et al., 2003). Snowmelt is the primary source of stream flow through the Teton canyons. Stream catchments draining from the eastern flank range in size from 1 to 64 km^2 (Foster et al., 2010).

This study focuses on Garnet Canyon, which is centrally located within the range and drains east toward the Jackson Hole basin (Fig. 1). The highest elevation in the Teton Range, the Grand Teton peak, defines the northwest corner of Garnet Canyon's drainage divide. Relative to other to catchments in the range, Garnet Canyon is a mid-sized catchment (drainage area = 9.8 km^2) with an average of 212 cm of precipitation per year, and a steeper longitudinal profile than catchments with drainage areas greater than 20 km^2 in the range (Foster et al., 2010). Water in the Garnet Canyon stream is provided by melting snow from the winter snowpack and ice from the Middle Teton Glacier. Cool temperatures and high elevations sustain the snowpack for most, if not all, of the summer.

The spatial distributions of stream and glacial erosion indicate that glacial incision effectively removed bedrock from the floor of Garnet Canyon during the last glacial advance (Tranel et al., 2015). Detrital apatite minerals from glacial sediments show erosion occurred near glacially flattened segments of Garnet Canyon and near the Pleistocene equilibrium line altitude of 2600 m (Pierce and Good, 1992; Foster et al., 2008; Tranel et al., 2011). Detrital apatite minerals from stream sediments imply a lack of erosion at the same locations where the glacier focused erosion (Tranel et al., 2011; Fig. 1b). Gaps in erosion indicate the potential for sediment storage in glaciated catchments that could pose problems for using detrital sediments to study a glaciated catchment.

3. Methods

Field data collection included cross section and sediment size measurements at seven sites in Garnet Canyon beginning at the western limit of the Meadows (below Spalding Falls) and continuing east to Cleft Falls (Fig. 1b and c) during the summer of 2008. The upstream-most observations came from two branches of the stream draining separate forks of the canyon before they merged into a single channel. Sites along the merged channel then progress downstream through the canyon between talus fan deposits sourced from the north and south facing canyon walls (Tranel et al., 2015). Glacial scour, hillslope deposition, and alluvial sediments characterize the valley floor within the study area (Figs. 1c and 2). Snow cover limited exposure of the stream channel in some areas. No suspended sediments were observed in the stream during data collection.

At each site, a tape measure was stretched across the width of the channel to include the bank and bankfull area above the active channel. Depth was recorded every half meter and at specific locations where boulders or islands emerged from the water. Bankfull measurements were excluded here due to the uncertainty of the exact bankfull position. Large boulder toes defining the bank could block flow, but because they were not continuous, could also allow water to flow past them beneath the talus material. The talus composition results in high porosity, which increases uncertainty in the bankfull extent and depth estimates.

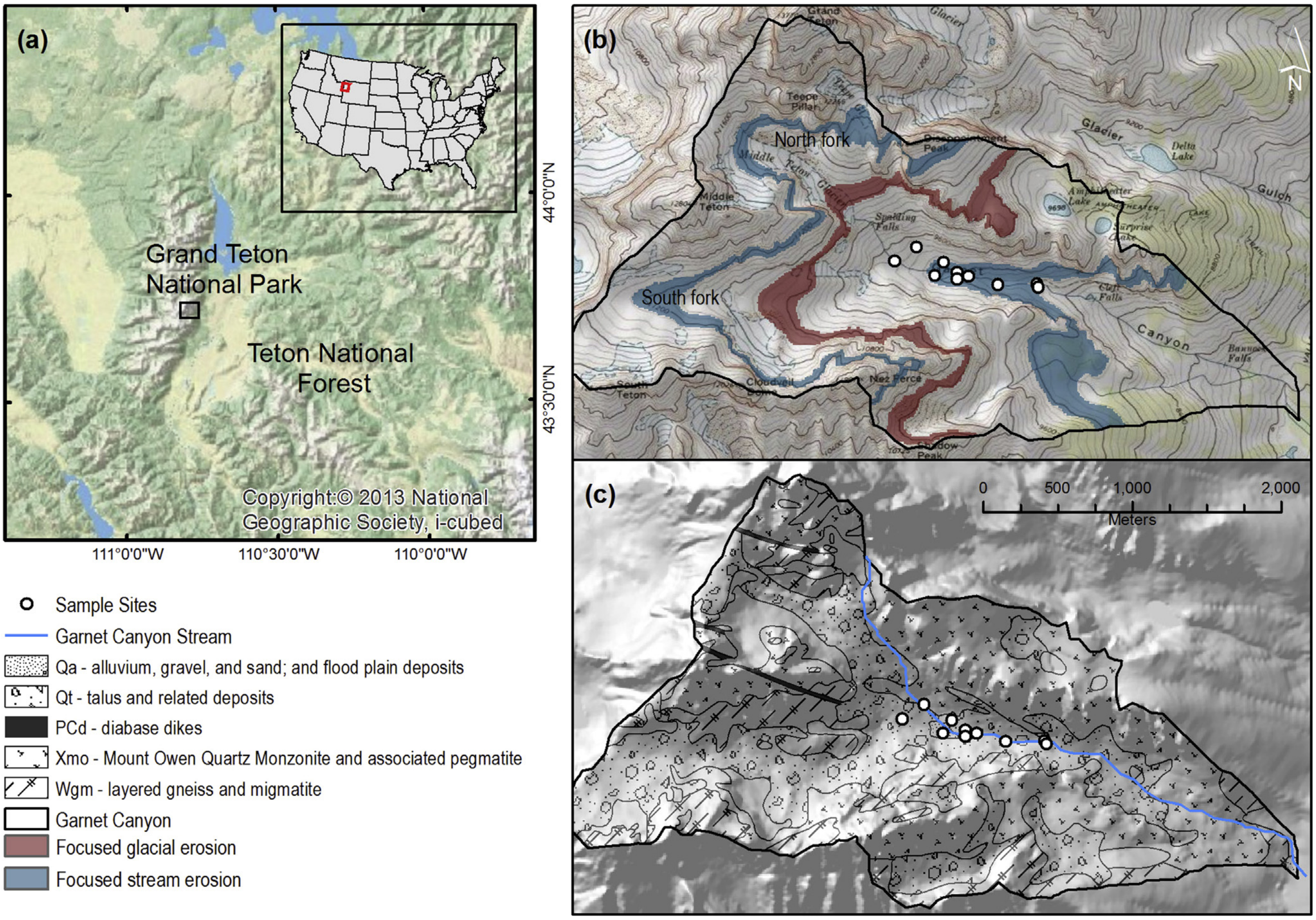

Figure 1. Garnet Canyon is located in the center of the Teton Range in northwestern Wyoming, USA. (a) The position of Garnet Canyon within the Teton Range is indicated by the black box. (b) Samples were collected in the center of the canyon along the valley floor. Blue and red shaded areas indicate focused erosion elevations observed in Tranel et al. (2011) based on the age-elevation relationship: $y = 29x + 2179$ where y is elevation and x is mineral age. The topographic basemap is the 1968 USGS Grand Teton 24,000 quadrangle map provided in ESRI's ArcMap program. Topographic contours are in feet. (c) The valley floor where samples were collected is covered with colluvial and alluvial sediments (geology map modified from Love et al., 1992). The blue Garnet Stream indicates the stream location defined by hydrology tools in ArcMap. The longitudinal profile line shown in Fig. 6 was taken from this line.

A random walk was applied to characterize surface bedload sediments at each cross section to quantify cumulative grain size distributions (GSD) (Wolman, 1954; Wohl et al., 1996). With each step across the stream, field assistants or I randomly selected a grain from the stream bed. The intermediate axis of each grain was measured with a granulometer. Large boulders that were too big to pick up were measured in place with a tape measure. We also noted if the bed was sandy when grains were too small to measure with the granulometer. Walks across the stream were continued until ~100 grains were observed (Wolman, 1954; Yager et al., 2012; Bacchi et al., 2014), shifting slightly upstream each time we needed to cross again. We took turns collecting because water temperatures were cold. Actual counts at a couple of sites fell short of 100 grains due to cold water and weather conditions. Counts at sites 8a and 8b were much lower than the standard 100 grains because in the field the data were collected counting the two branches as parts of a single channel divided by an island. Upon assessing results, the branches were separated into separate cross sections to account for cross section classification differences that were observed between the two branches. The result, however, is limited compared the number of observations required for better statistical assessments of grain size descriptions (Rice and Church, 1996). Finally, graphic means and standard deviations were calculated with equations given by Folk and Ward (1957).

To quantitatively assess the similarity in stream characteristics to other mountain channel reaches, roughness was calculated for each stream section using mean flow depth divided by D_{95}, both measured in meters. With our field data, we also estimated shear stress conditions and quantified how those conditions change with different mineral densities. The boundary shear stress (τ) was estimated for simple, steady flow using the equation (Zimmermann and Church, 2001; Torizzo and Pitlick, 2004; Wohl and Wilcox, 2005):

$$\tau = p_{\mathrm{w}} gRS \tag{1}$$

where the hydraulic radius, R, was calculated from the area divided by the wetted perimeter of the active stream observed in the field; slope, S, was measured within the stream along a minimum length of 3 channel widths. The density of water, p_{w}, is 1000 kg/m^3 and g is the acceleration due to gravity (9.81 m/s^2). Critical shear stress (τ_{c}) was calculated based on the equation listed in Wohl and Wilcox (2005) where:

$$\tau_{\mathrm{c}} = t^{*}_{\mathrm{c}}(p_{\mathrm{s}} - p_{\mathrm{w}})gD \tag{2}$$

The standard density value used for sediment is 2650 kg/m^3, based on the density of quartz. The critical shear stress was also

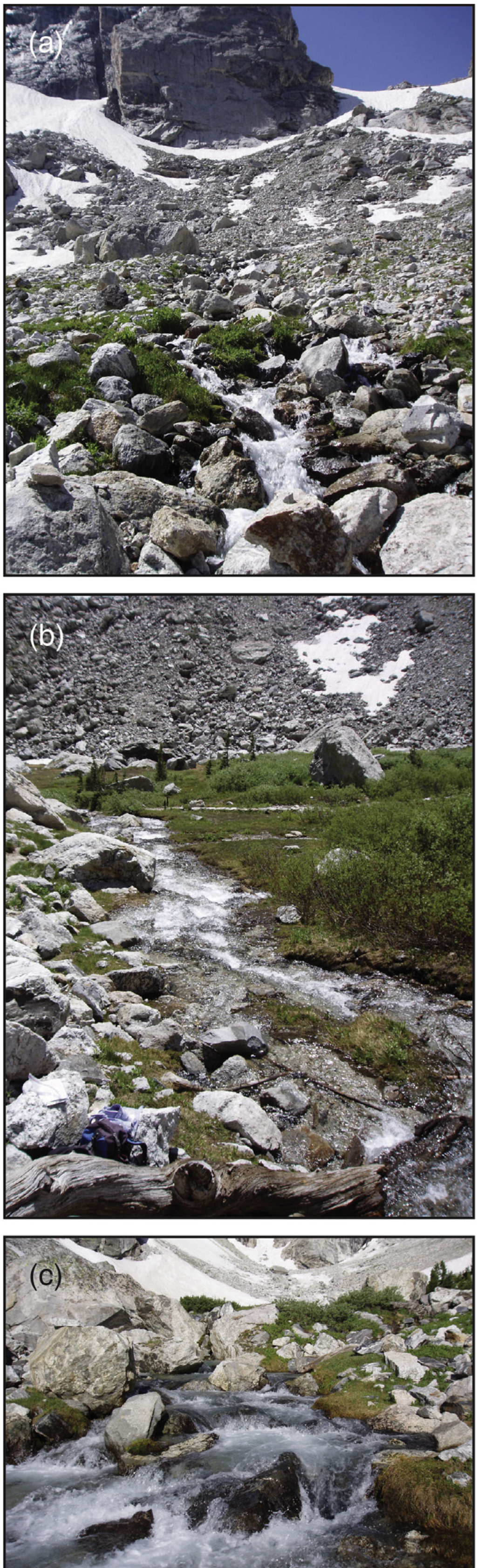

Figure 2. Photos illustrating three classes of streams based on the relative position over or near talus deposits. (a) Colluvial: the stream flows directly over talus deposits. (b) Coupled-one bank: the stream flows directly adjacent or within a few meters of the toe of the talus fan on one bank but the other bank tapers off into a flat alluvial area with randomly scattered rock debris. (c) Coupled-both banks: the stream flows between the toe of a talus fan on either bank. Distance to the talus fan toe is only a few meters on either side.

calculated using the densities of apatite minerals (3190 kg/m^3) and zircon minerals (4600 kg/m^3). Apatite and zircon densities were selected for this study because these mineral systems have been increasingly used in recent research to trace spatial patterns of erosion based on low temperature bedrock and sediment thermochronometry (Tranel et al., 2011; Enkelman and Ehlers, 2015; Riebe et al., 2015). Grain size (D) was varied to assess the differences in shear stress to move different size grains that might be sampled in detrital studies. A D-value of 2 mm was used because it represents the sediment sizes typically scooped out of a streambed for detrital cosmogenic nuclide or thermochronometry studies (Riebe et al., 2015; Lukens et al., 2016). The D_{50} value was used for D in Eq. (2) because it represents the standard grain size used in river or stream critical shear stress calculations (Wohl and Wilcox, 2005). Lastly, D_{84} was also used in Eq. (2) to assess critical shear stress as recommended by recent investigations of steep mountain stream systems (Recking, 2012; Recking et al., 2012). Shield's parameter for the critical shear stress equation was calculated based on the channel slope as recommended by authors studying streams in steep mountain landscapes (Recking, 2012; Bacchi et al., 2014):

$$T^{*}{}_{c} = 0.15 \times S^{0.275} \tag{3}$$

In 2011 we returned to the field and repeated grain size measurements at 4 sites. Differences between GSD were observed, however, the differences did not change the assigned classifications or roughness. In a study comparing different methods of random selection in stream sediments, Wohl et al. (1996) found that there are statistically significant differences in grain size distribution data collected by novice or expert field assistants. The data from the second field season in this study (2011) are excluded from the figures because the assistants randomly selecting sediments in the current study were different between 2008 and 2011 field seasons and fewer sites were observed the second year.

4. Field sites

Much of the Garnet Canyon stream channel fits the description of a sink colluvial channel described by Brardinoni and Hassan (2006) because it receives colluvium input laterally and the valley is too narrow to prevent the colluvium from entering the stream. The following results summarize the coupled relationships observed in Garnet Canyon (Fig. 2) and the cross section sites (Table 1 and Fig. 3) described as compared to classifications by Montgomery and Buffington (1997) and Halwas and Church (2002). Upstream tributary sections of the channel are colluvial, and water flows directly over talus deposits or glacial drift. The banks are composed of very large boulders and cobbles. This type of flow is most likely to occur on the upstream-most sections of the stream and at higher elevations where less vegetation growth occurs and surfaces may only be snow free for a short period over the summer. The branch of the stream exiting the north fork of the canyon initially flows over glacially polished bedrock creating a large waterfall (Spalding Falls) and then between colluvium and possibly some glacial till. Water flow through boulders creates a cascade channel (site 3) that looks similar to cascades reaches (Montgomery and Buffington, 1997; Halwas and Church, 2002). The stream from the south fork follows a complicated path below

Table 1
Stream reach characteristics and classifications.

Site	Elevation (m)	Fractional slope	Mean flow depth (m)	Roughness	Montgomery & Buffington classification	Halwas & Church classification	Downstream distance (km)
3-2008	2871	0.17	0.20	0.50	Cascade (step-pool)[a]	Boulder cascade (rapid,chute)	1.5
4-2008	2822	0.12	0.29	0.24	Step-pool	Chute	1.7
6-2008	2813	0.09	0.29	0.68	Step-pool	Rapid, chute (riffle)	0.5
8a-2008[b]	2810	0.08	0.11	0.55	Plane-bed (step-pool)	Riffle	0.7
8b-2008[b]		0.08	0.16	1.17	Cascade	Cascade	1.9
8-2011		*0.08*	*0.08*	*1.20*			
2-2008	2786	0.04	0.19	0.53	Cascade (plane bed)	Rapid, chute (riffle)	2
2-2011		*0.04*	*0.21*	*0.19*			
7-2008	2755	0.04	0.63	0.80	Step-pool (plane bed)	Rapid (riffle)	2.3
7-2011		*0.04*	*0.33*	*0.32*			
1a-2008	2731	0.06	0.22	0.08	Pool-riffle	Rapid, chute (glide)	2.7
1a-2011		*0.06*	*0.20*	*1.53*			
1b-2008		0.06	0.07	1.57	Cascade	Boulder cascade	
1b-2011		*0.06*	*0.06*	*7.81*			

Italicized rows summarize results from the second field season in 2011, excluded from figures and shear stress analyses.
[a] Classification in parentheses indicates mismatch between qualitative and quantitative descriptions.
[b] SM08-08a and SM08-08b reaches are on north and south branches of the stream respectively.

Figure 3. Photos from the observed cross section sites and corresponding position along the length of the stream. The aerial image was provided courtesy of Grand Teton National Park.

snowfields and beneath talus debris until it appears continuously in the Meadows, also flowing over talus, as a step-pool reach at site 6.

In a second coupled relationship observed in Garnet Canyon, the stream flows adjacent to a talus fan toe on one bank. The opposite bank is adjacent to a vegetated alluvial surface (sites 8a and 8b). The talus toe usually contains large boulder and cobble sized sediments. The flat surface above the opposite bank is vegetated with grasses, flowers, and small bushes. Additional water sometimes flows across the surface when snowmelt is high or after heavy rainfall. Randomly positioned, large boulders also frequently occur. Site 4 is the first section with the single coupled bank observed downstream from site 3, along the north edge of the flat meadow. The talus is sourced from the nearby south-facing canyon wall. Cross-section 4 resembles Montgomery and Buffington's (1997) description of a step pool reach. Farther downstream, at the eastern-most edge of the meadow, the stream transitions to a plane bed reach (site 8b) just before joining with the channel from the south fork (Fig. 3). Site 8a is directly south of site 8b, and downstream from site 6 on the south branch of the stream. It also fits the description of a plane bed reach as it flows adjacent to the talus toe to the south of the channel.

In four of our cross-section sites, the stream flows between two talus fan toes. These segments are coupled to hillslope talus along both banks, although the stream frequently has a narrow, vegetated bank between the channel and the talus toe. Three of these sites are also located where the canyon walls narrow. The canyon narrows downstream of the Meadows area where the north and south branches of the stream merge. Rockfalls sourced from steep valley walls deposit debris that stretches across the width of the canyon floor (Tranel et al., 2015). Because the stream interacts more closely with coarse talus debris again, the channel resembles a cascade reach at site 2. At the next location downstream, patches of grass cover a narrow strip of bank between the channel and the toes of talus debris on either side of the stream. Within the channel at site 7, the stream features a step-pool reach. The stream at the final site divides to flow north and south around a patch of small bushes and boulders. One side is characteristic of cascade flow (site 1a) and the other side is characteristic of pool-riffle flow (site 1b).

Downstream from site 1, the stream washes over exposed bedrock again as it flows over Cleft Falls. Observations ended at this point because it was a steep scramble to reach the channel below the waterfall. Additionally, the falls mark the close proximity to the end of the narrow segment of the canyon. Downstream of this point, the canyon widens again and the stream approaches the steep eastern front of the mountain range.

5. Results

To assess spatial variability of channel morphology in Garnet Canyon, we compared slope, roughness and grain sizes between observation sites. The slope decreases consistently from upstream to downstream until the lowest site, where it increases again by 2% (Table 1). Despite the decreasing trend in slope, slope-roughness relationships are not distinct for the different classifications in Garnet Canyon (Fig. 4). Stream sediments are poorly sorted gravels in each cross section except for reach 1a, which is moderately sorted (Table 2). Graphs displaying the cumulative percent finer GSD indicate that the largest grains exist in an upstream-most reach (site 6) and the smallest grains appear at the lowest downstream site (1a; Fig. 5). Most reaches contain similar GSD unrelated to the Montgomery and Buffington (1997) classification. Other sites show no variation in GSD with distance downstream (Fig. 6). A Kolmogorov-Smirnov test comparing the cumulative curves for each site to an average curve created from all of the sites confirmed that all sites excluding 1a and 6 are not statistically different from each other (Table 3). Tables 2 and 3 also summarize findings for the GSD observed in 2011. Statistical comparison between years indicated that the GSD statistically were not the same in the two observed years. I excluded observations from 2011 from further calculations due to concerns about differences in random selection and observations as explained earlier in the methods.

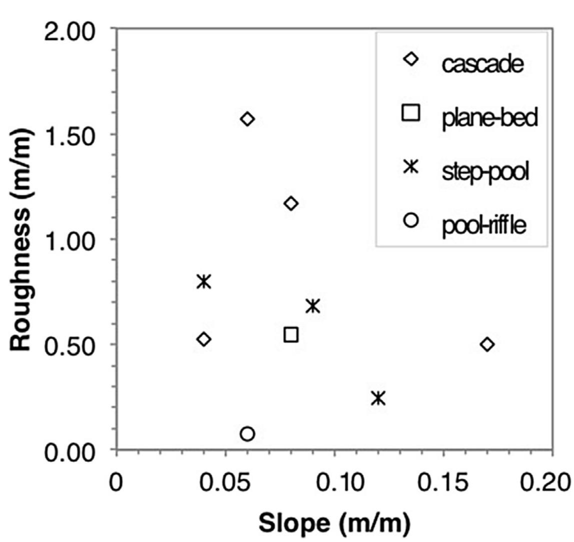

Figure 4. Slope and roughness results show scatter across reach classifications.

Table 2
Sediment size distributions observed in Garnet Canyon.

Site	Clasts	Grain size (mm)							Sorting
		D_5	D_{16}	D_{50}	D_{84}	D_{95}	Mean	Standard deviation	
3-2008	102	4.2	6.1	18	61	100	18.8	2.9	Poorly
4-2008	110	2.5	4	16	43	70	14.0	3.0	Poorly
6-2008	98	4.2	11	70	130	200	46.4	3.3	Poorly
8a-2008	44	2	9	21	50	60	21.1	2.6	Poorly
8b-2008	39	2	2.8	20	60	190	15.0	4.3	Very poorly
8-2008	*100*	*2*	*2*	*10*	*50*	*100*	*10.0*	*4.0*	Very poorly
2-2008	99	2	5	29	60	100	20.6	3.4	Poorly
2-2011	*101*	*2*	*2*	*2*	*27*	*40*	*4.8*	*3.0*	Poorly
7-2008	97	5.1	8	14	28	500	14.6	2.7	Poorly
7-2011	*95*	*2*	*10*	*18*	*39*	*105*	*19.1*	*2.6*	Poorly
1a-2008	97	2	5.2	9	13	17	8.5	1.7	Moderately
1a-2011	*50*	*3*	*5*	*11*	*50*	*305*	*14.0*	*3.6*	Poorly
8b-2008	102	3.9	7.1	17	65	110	19.9	2.9	Poorly
8b-2011	*30*	*7*	*12*	*100*	*290*	*500*	*70.3*	*4.2*	Very poorly

Italicized rows summarize results from the second field season in 2011, excluded from figures and shear stress analyses.

Shear stress conditions were evaluated based on field observations from the 2008 field season to assess transport potential. The observed flows in the active channels in Garnet Canyon were sufficient to move quartz density sediments in all sections with 2 mm or D_{50} sized grains. The calculations indicate that flow was also sufficient to move D_{84} grains in roughly half of the stream sections. While we were in the field, the water ran clear and bedload movement was not obvious. This discrepancy between our flow calculations and lack of bedload movement observations while in the field may result from a number of uncertainties. One consideration is a bias while selecting grains for the GSD. Water was cold and flowing fast, so our efforts were to limit time spent in the stream and may have caused a bias toward collecting smaller grains. Future improvements to these observations could include multiple sampling techniques to assess differences between them. In ideal conditions, we would use sediment traps to observe actual sediments transported, however, we chose a more discrete method due to our location within a national park in an area frequented by visitors. Additional uncertainties are related to the representation of the boundary flow equation as applied to a steep mountain stream, potential sediment shielding related to the poorly sorted nature of the sediments, or daily fluctuations in flow related to the timing of snowmelt.

Changing the density did influence the results determining which grain sizes could be transported in the active flow. The boundary shear stress was sufficient to transport quartz density sand and D_{50} grains through all channel sections. It was also sufficient to move D_{84} grains through 5 of the 9 sections (Table 4 and Fig. 7). Changing the density to equal that of apatite or zircon, the grain sizes the stream efficiently transported decreased. Site 1a would not allow D_{50} apatite density grains through, and three sites could not transport D_{50} zircons. The boundary shear stresses were sufficient to transport D_{84} apatite or zircon density materials in only two observed sections.

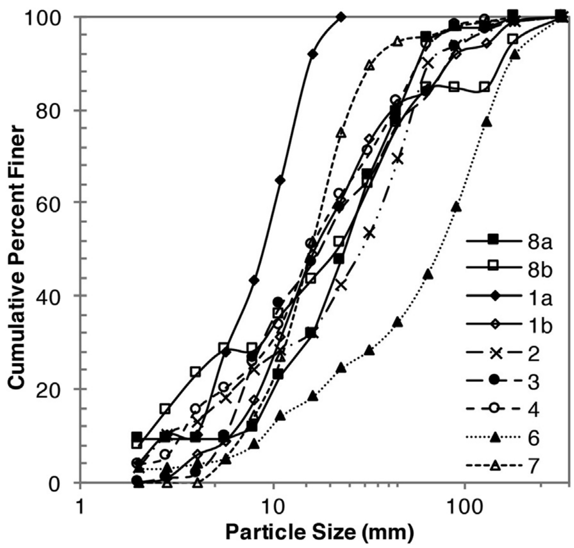

Figure 5. Sediment size cumulative percent curves for each stream section in Garnet Canyon.

6. Discussion

6.1. Talus-stream morphology

The Garnet Canyon stream flows over, around, and between talus as it makes its way out of the narrow, glaciated canyon in the center of the Teton Range. Along the reach where observations were collected, the channel is coupled with hillslope deposits on at least one bank of the stream. When the valley is snow free, the only path for a hiker through the canyon requires climbing over large boulders or walking through the stream channel. Although the stream is coupled with the hillslopes, it cannot be classified as purely colluvial because evidence also indicates that alluvial processes act within the stream channel. The channel is clearly defined along the length of the stream. Poorly sorted sediments within the stream channel are smaller in size and distributed differently than the surrounding talus sediments. Channel slopes at our sites were taken from some of the lowest slope areas along the entire longitudinal profile of the stream (Fig. 6), and therefore, also lower than would be expected in typical colluvial stream reaches.

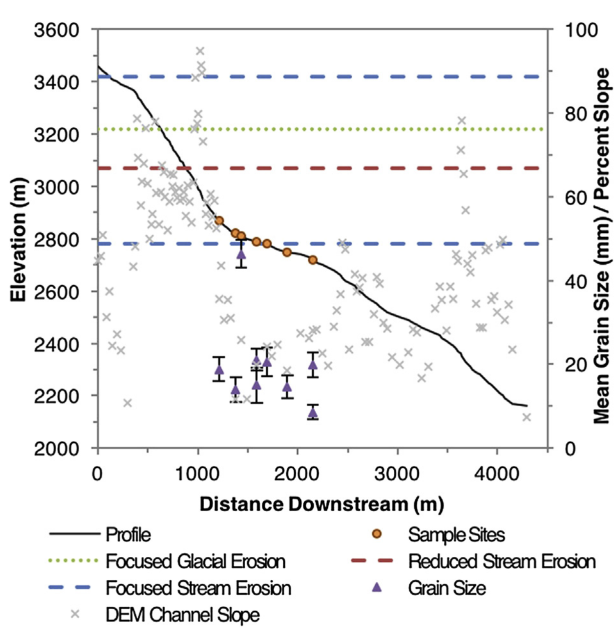

Figure 6. Longitudinal stream profile indicating sample location sites. The profile begins in the north fork of Garnet Canyon where the Middle Teton glacier still covers the surface of the valley floor for a distance upstream of Spalding Falls (Garnet Stream in Fig. 1b). The slope values in this graph were derived along the stream channel from a 10 m digital elevation model (DEM). Mean grain sizes plotted at their position in place along the longitudinal stream profile. Error bars represent standard deviation calculated using methods of Folk and Ward (1957). Dashed lines indicate focused or limited stream or glacial erosion as indicated by Tranel et al. (2011).

While not entirely colluvial along its length, the Garnet Canyon stream is also not in equilibrium. It demonstrates some downstream change between stream classes that is expected in an equilibrium system, but the trend is repeated. Hillslope coupling could play one part in the repeated sequence of channel characteristics because the hillslopes can deposit large cobbles within the stream channel that are locally redistributed to create barriers in the step pools or provide an abundance of smaller gravels and cobbles that are redistributed and deposited downstream (Montgomery and Buffington, 1997; Church, 2010; Table 1 and Fig. 3). The expected evolution of channel reach characteristics defined by Montgomery and Buffington (1997) include a progression from cascades at high elevations to plane beds at low elevations. In steep mountain streams, however, external processes may disrupt or randomize downstream channel morphology patterns (Zimmermann and Church, 2001; Halwas and Church, 2002; Chin and Wohl, 2005). Because glacial geomorphology influences both hillslope and stream processes, Brardinoni and Hassan (2007) characterized channel reaches in a glaciated landscape. They observed repeated patterns strongly tied to hanging valleys. Garnet Canyon is different from the valleys that they studied, in that it is a relatively simple and straight canyon. The north and south forks are tributaries, running parallel to each other and merging to form the trunk canyon. Downstream from the fork junction, there are no other significant tributaries or hanging valleys. Slope and talus deposits are the only conditions that can influence the channel morphology within that straight and narrow segment of the Garnet Canyon stream channel. Slope in particular is influential in defining hydraulic conditions in glaciated catchments (Brardinoni and Hassan, 2006).

Additional evidence from our field observations underscores the disequilibrium of the Garnet Canyon stream. Slopes, roughness, and grain size distributions are not distinct between sites in Garnet Canyon and values do not fit within expected published values from other steep mountain systems (Howard et al., 1994; Montgomery and Buffington, 1997; Halwas and Church, 2002). The similarity in grain sizes may reflect a similar sized population of sediments input into the stream and moving through the channel during high flow events (Zimmermann and Church, 2001). Sediment distributions likely represent a combination of random deposition from hillslopes and stream capacity to transport particular sizes farther downstream (Whittaker, 1987; Wetzel, 1994; Chin and Wohl, 2005). Improvements to this study to better understand sediment transport efficiency and coupling between hillslope and stream processes include longer-term observations to trace or collect sediments and detailed descriptions of colluvial sediments compared to alluvial sediments. Individual grain shape characteristics may identify links to glacial and hillslope sediments (Lukas et al., 2013). Collecting grains with sediment traps would verify the size of material that can move during snowmelt flows and also identify the sediment sizes transported during high flow events. Additional quantitative observations of overland flow during storm events could characterize the sediment sizes that can be transported to the stream channel from talus sources as a result of overland or through talus flow. Time constraints in the field during the original study limited the data and observations that could be collected in the field.

Table 3
Kolmogorov-Smirnov statistical comparisons.

Site	Comparison site	Grain count	D*	Significance	Accept or reject null hypothesis
Comparison between GSD at each site and average GSD of all sites. The null hypothesis is that the distributions are not different.					
3-2008	Average	106, 104	0.07	>10	Accept
4-2008	Average	111, 104	0.12	>10	Accept
6-2008	Average	93, 104	0.4	<1	Reject
8-2008 combined	Average	86, 104	0.1	>10	Accept
2-2008	Average	102, 104	0.15	>10	Accept
7-2008	Average	102, 104	0.23	1–5	Accept
1a-2008	Average	97, 104	0.45	<1	Reject
1b-2008	Average	102, 104	0.07	>10	Accept
Comparison between GSD observed in 2008 and 2011. The null hypothesis is that the distributions are not different.					
8-2008 combined	*8-2011*	86, 93	0.46	<1	Reject
2-2008	*2-2011*	102, 102	0.21	<1	Reject
7-2008	*7-2011*	102, 112	0.25	<1	Reject
1a-2008	*1a-2011*	97,81	0.46	<1	Reject
1b-2008	*1b-2011*	102,81	0.57	<1	Reject

Italicized rows summarize results from the second field season in 2011, excluded from figures and shear stress analyses.

Table 4
Summary of hydraulic conditions.

Site	Width (m)	R (m)	Shield's value	τ	τ_c medium sand (2 mm)			τ_c D_{50}			τ_c D_{84}		
					Quartz	Apatite	Zircon	Quartz	Apatite	Zircon	Quartz	Apatite	Zircon
3-2008	3.4	0.21	0.0921	350	2.98	3.96	6.51	26.8	35.6	58.6	91.0	121	199
4-2008	2.2	0.04	0.0837	49	2.71	3.60	5.91	21.7	28.8	47.3	*58.3*	*77.3*	*127*
6-2008	3.5	0.15	0.0774	132	2.50	3.32	5.46	87.7	116	*191*	*163*	*216*	*355*
8a-2008	4.5	0.09	0.0749	71	2.42	3.22	5.29	25.5	33.8	55.5	60.6	*80.4*	*132*
8b-2008	3.2	0.10	0.0749	78	2.42	3.22	5.29	24.2	32.2	52.9	72.7	*96.5*	*159*
2-2008	5.34	0.16	0.0619	63	2.00	2.66	4.37	29.1	38.6	*63.4*	60.1	*79.8*	*131*
7-2008	2.5	0.22	0.0619	86	2.00	2.66	4.37	14.0	18.6	30.6	28.1	37.2	61.2
1a-2008	1.95	0.02	0.0692	12	2.24	2.97	4.89	10.1	*13.4*	*22.0*	*14.6*	*19.3*	*31.8*
1b-2008	1.63	0.10	0.0692	59	2.24	2.97	4.89	19.0	25.3	41.5	*72.8*	*96.6*	*159*

R is hydraulic radius. τ_c is critical shear stress. τ is boundary shear stress. Density of water was 1000 kg/m^3. Density of quartz sediment used was 2650 kg/m^3. Density of apatite used was 3190 kg/m^3. Density of zircon used was 4600 kg/m^3. Italicized values are critical shear stresses that were greater than the boundary shear stress.

6.2. Sediment characteristics related to transport and detrital studies

The differences in channel morphology observed along the study length of the Garnet Canyon stream did not create significant differences in the stream's capability to transport detrital sediments used for erosion rates or spatial indicators. The observed snowmelt discharges were sufficient to transport D_{50} sizes observed in the active stream sections. Talus deposits and glacial scour did not decrease slope sufficiently to reduce the capacity for sediment transport based on the observed grain size distributions and observed flow conditions in this study. Some mountain catchments experience severe overdeepening that significantly lowers the channel slope and traps sediments upstream (Dühnforth et al., 2008; Straumann and Korup, 2009). The relatively small catchment area of Garnet Canyon in comparison to other canyons in the Teton Range limited glacial erosion (Foster et al., 2010). The small size consequently prevented the glaciers from reducing slope enough to influence sediment transport efficiently at these study sites, despite positioning observation locations within the flattest upstream segment of the canyon (Figs. 1b and 6).

Sand sized sediments, most frequently collected for detrital studies, were only found in small patches at our cross section sites. The rapid flow observed in the Garnet Canyon stream suggests that most sand is quickly carried downstream upon entering the active stream channel. Sand is easily entrained from banks and streambeds and transported downstream during moderate flows in steep channels because the water velocity is greater than the settling velocity for sand (Warburton, 1992; Garzanti et al., 2008; Church, 2010; Yager et al., 2012). Some uncertainty may be associated with the limited study area size, in addition to the uncertainty associated with the general complexity of steep glaciated mountain stream systems (Brardinoni et al., 2015). Additionally, equations applied to mountain streams may not accurately represent the bedload stability or transport dynamics in these complex systems (Yager et al., 2012).

Additional complexity in the Garnet Canyon stream system is due to the heterogeneity of bedrock mineralogy. While the flow in the Garnet Canyon stream was sufficient to transport D_{50} and some D_{84} grains, even when heavy zircon and apatite densities are applied, actively moving grains were not observed in the channel. Garnet Canyon is primarily composed of the Mount Owen quartz monzonite and undifferentiated layered gneiss in Garnet Canyon. These rocks are primarily silicate minerals within the typical density applied to shear stress equations for hydraulic conditions (70–75 wt.% SiO_2; Frost et al., 2006). Garnet Canyon also contains several unique, high-density rock units randomly distributed throughout the catchment (Fig. 1c). Anomalously high density grains in the stream channel can include sediments from mafic dikes that intrude the rock across Garnet Canyon or amphibolite that exists in very large, discrete blocks within the layered gneiss (Frost et al., 2006). Higher density minerals composing those rock units and gravels or cobbles potentially sourced from these units are similar to apatite density applied in this study (example densities: amphibole = 2850–3570 kg/m^3; biotite = 2700–3500 kg/m^3; garnet = 3100–4200 kg/m^3; Nesse, 2017). Hillslope failures and

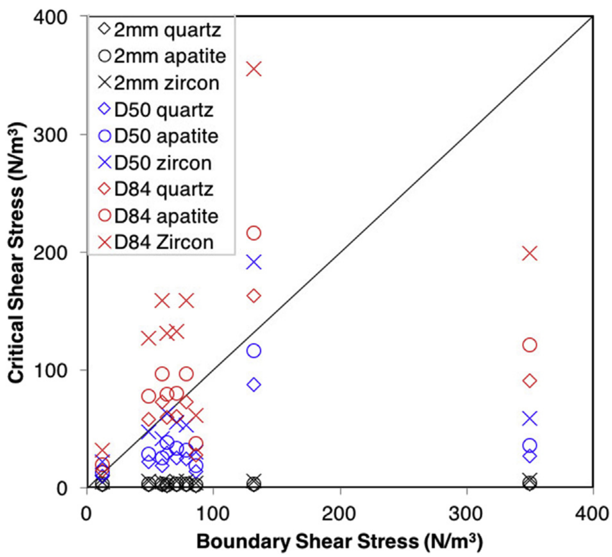

Figure 7. Comparison between boundary shear stress and critical shear stress at each channel cross section for each mineral density evaluated in this study.

glacial debris include these materials when they are present, and therefore create higher density sediments supplied to the stream channel.

Gravels composed of mafic dike or amphibolite could armor the bed surface with high density sediments resulting in a stable bed despite water flows sufficient to move away sediments based on the GSD. Immobile sediments in stream channels can prevent movement of bed sediments (Bathurst, 1987; Church, 2010; Bacchi et al., 2014). The strong link with hillslope processes indicates that some dense gravels or cobbles could simply be dropped into the stream channel to hold lower density sediments in place. Higher density rocks would also be moved during higher flow conditions than those observed in this study. In addition to a bedload armor, jams can develop with large grains blocking smaller grains from moving. Once formed, the stream requires more force to move grains past the jam (Church, 2010). Given potential difference in density, grains may not need to be much larger than surrounding grains to start resisting entrainment and blocking flow. If a portion of the bedload armor consists of high density rocks, the flow required to mobilize nearby or covered, less dense grains would further increase. Few rocks would contain enough heavy minerals that the gravels would have densities as high as zircons, however, apatite density may approximately represent the amphibolite or mafic dike rock densities in shear stress equations estimating the conditions required to entrain sediments. From these results, we see that density matters in transport models in steep mountains streams. Future work should consider lithology and mineralogy, as well as mineral shape, to assess transport and settling velocity potential in the stream channel (Garzanti et al., 2008; Lukas et al., 2013).

The evaluation of critical shear stress in this study indicates that regardless of density, sands and small gravels are easily transported throughout the Garnet Canyon stream. The efficiency of the system is verified by the presence of many sand grains sourced from high elevations observed in detrital apatite minerals (Tranel et al., 2011). A map identifying the source of apatite minerals based on the age-elevation relationship illustrates the distance those sand grains traveled and the talus cover that offered potential barriers extending across the canyon (Fig. 1b). The longitudinal profile also highlights some slope reduction 1–2 km downstream from the headwaters (Fig. 6). Detrital apatite minerals also indicate some stream focused erosion where slope decreases between the Meadows and Cleft Falls (Tranel et al., 2011). Detrital sediments from within this area seemingly armored by talus may indicate that sediment thickness overlying bedrock is not great and the abundant debris that can move through the channel can also cause some scour on the underlying bedrock surface. A comparison of detrital gravels would be valuable to further test the results here and the efficiency of sediment transport throughout Garnet Canyon.

7. Conclusion

Along the length of the Garnet Canyon stream, the channel morphology changes as a result of how the stream couples with talus deposits and how sediments organize in the channel to direct water flow. Despite the variations in channel morphology, grain size distributions were similar, and flow conditions were sufficient to transport gravels throughout the sites in this study. Transient landscapes can be complicated to study with detrital mineral geochronology or thermochronology. Recently glaciated catchments in particular are challenging due to the strong coupling between glacial, hillslope, and fluvial features. Garnet Canyon is certainly both complex and transient, however results in this study indicate that concerns for storage upstream due to accumulation where glacier scour reduced slopes or hillslopes deposited talus are not necessary for evaluation of detrital sediments. The next part of the story to consider in detrital studies of a coupled system like Garnet Canyon, is the genetic history of the sediments, whether they are derived from glacial, hillslope, or fluvial erosion.

The summary of a single canyon limits the ability to identify trends that apply to all glaciated catchments, but it provides a starting point to discuss future work on streams at various scales within glaciated catchments and composed of bedrock with density variations. Garnet Canyon is a mid-sized catchment, and the importance of mid- to small-sized features tends to be excluded from regional models. Works by Brardinoni and Hassan (2006, 2007) began assessment of relationships between glacial, hillslope, and stream conditions in glaciated catchments. Because climate, tectonics, and lithology all influence the coupled glacial-colluvial-fluvial systems, many more observations across many mountain ranges are needed to better understand transport efficiency and channel evolution in alpine landscapes. Additionally, the bedrock composition is a condition that may need more frequent consideration in certain landscapes with mixed lithologies.

Acknowledgements

I thank Sally Morgan, Amber Ritchie and Audrey Happel for field assistance and Emily Zawacki for sieving sediments. The University Research Grant program at Illinois State University and the Graduate Research Development Program Grant at Virginia Tech provided funding for field support.

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