The Effects of Variable Precipitation on Discharge and Sediment Transport in Streams in the Teton Mountain Range, Wyoming, USA

Amber J. Ritchie[1], Lisa Tranel[2]

Geography-Geology Department, Illinois State University, Normal, IL 61761, USA

Abstract— *Discharge and sediment distributions control the efficiency of sediment transport and incision into bedrock units in active stream channels. The efficiency of stream erosion is an important factor influencing the evolution of mountain landscapes. Variations in yearly precipitation affect the timing of snowmelt, and therefore, the water availability for discharge in high elevation streams. This study explores how differences in annual precipitation can impact alpine stream erosion. Water discharge, bed load sediments, and suspended solids were observed for major streams draining watershed areas between 10 km² and 43 km² in the Teton Mountain Range in northwestern Wyoming, USA. The maximum sediment sizes capable of being moved through the stream channels at late summer flow conditions were determined using basal shear stress and critical shear stress calculations. Annual precipitation data over 2 years was compared with sediment transport conditions to compare how precipitation impacted erosion. Erosion proved to be effective in both high and low precipitation conditions; however higher precipitation resulted in prolonged snowmelt, higher discharge, greater sediment transport, and therefore higher erosional efficiency.*

Keywords— Teton Mountains, alpine streams, discharge, sediment transport.

I. INTRODUCTION

Streams play an important role in the evolution of many mountain landscapes by acting as transport mechanisms to move sediments from high to low elevations (Whipple et al., 2000; Kirby and Whipple, 2001; Tomkin et al., 2003).River incision into bedrock is a key erosion process controlling the rate of landscape responses to change in rock uplift rate and climate in mountainous areas (Howard, 1998; Whipple et al, 2000).The efficiency of stream erosion is influenced by the availability of water and sediments in the channel, which provide energy and tools respectively to drive transportation and incision. In alpine systems, water sources include precipitation, snowmelt and glacier melt, all of which are sensitive to small changes in climate conditions (Wendel, 2015). Stream sediments are sourced from the active channel, colluvial deposits below steep hillslopes, and glacial moraines or till (Wohl, 2005).

Precipitation is an important factor influencing stream discharge and erosion. Rainfall runs overland to enter the stream channel; and snowmelt accumulates over winter and melts throughout the summer at high elevations in alpine mountains. Intense storms or extreme temperatures driving rapid snowmelt can cause downstream flooding. When streams reach high flow or flood-like conditions, channel morphology undergoes its greatest changes (Leopold and Maddock, 1953). At high discharge, streams are more likely to incise bedrock, creating deeper flow and steeper slopes (Park 1977).

Sediment sizes and volumes carried by streams are also indicators of how much erosion occurs within watersheds (Tomkin et al., 2003). Sediments carried by streams act as tools to abrade streambeds and cause incision (Sklar and Dietrich, 2001). If sediments accumulate in a thick layer in the streambed, erosion is focused further upstream (Wohl, 1998). As rock erodes from its source and is carried downstream, 30% of sediment eroded will be transported through the length of the stream system to the mouth (Walling, 1983). Maximum erosion occurs when bedrock is only partially exposed under a coarse-grained supply of sediment. Fine-grained sediments in streams (clay, silt, and sand sized particles) abrade channels less effectively because they are mostly transported in suspension (Sklar and Dietrich, 2001). Larger clasts, including gravel, pebbles, cobbles, and boulders, therefore, play a more important role in stream abrasion and incision.

Small mountain streams receive comparatively less attention than larger, alluvial rivers due to difficulties in accessing and monitoring these systems (Montgomery and Gran, 2001). To aid in understanding how small stream systems effectively erode the landscape, we investigate summer stream flow in the Teton Range in northwestern Wyoming over two years with different precipitation records. The Teton Mountains have a distinct landscape influenced by glacial, fluvial, and mass wasting erosional forces. The result of these forces is a steeply sloping topography lacking vegetation on the mountain slopes composed of resistant bedrock. Streams were previously studied to understand the rates and patterns of erosion in Garnet and

Cascade Canyons (Tranel et al., 2011 and 2015). Mass wasting provided many sediments from high elevation hillslopes; and glacial erosion caused limits to erosion (Tranel et al., 2011). Although the previous studies used stream sediments, they revealed more information about the importance of hillslope and glacial erosion than the efficiency of modern streams in the Teton Range.

In this project, we assess how efficiently the streams erode their respective canyons. To do this, we measured stream discharges in different watersheds in the Teton Range to examine sediment transport, and how it is controlled by yearly variations in precipitation. Bed load sediment sizes and suspended solids were analyzed in four watersheds of the Teton Range to determine how catchments vary in the materials they transport. By researching how snowmelt influences sediment transport in the Teton Range, we hope to better understand how the stream system contributes to the evolution of this complex landscape.

II. STUDY AREA

Distinct and jagged peaks make up the 64 km long (north-south) and 24 km wide (east-west) Teton Mountain Range in northwestern Wyoming (Fig. 1). The geologic history the range includes Cretaceous and Neogene faulting, Quaternary glaciation, and volcanism associated with the Yellowstone volcanic high (Love et al., 2003; Fig. 1). Teton Mountain uplift began with the Laramide orogeny (55-80 Ma) and involved both faulting within basement rock and folding Paleozoic sedimentary units (Craddock et al., 1988; Roberts and Burbank, 1993). The younger Teton Normal Fault initiated 13-24 Ma (Love et al., 2003). Quaternary movement averages 1.3 mm/yr (Pickering White et al., 2009).

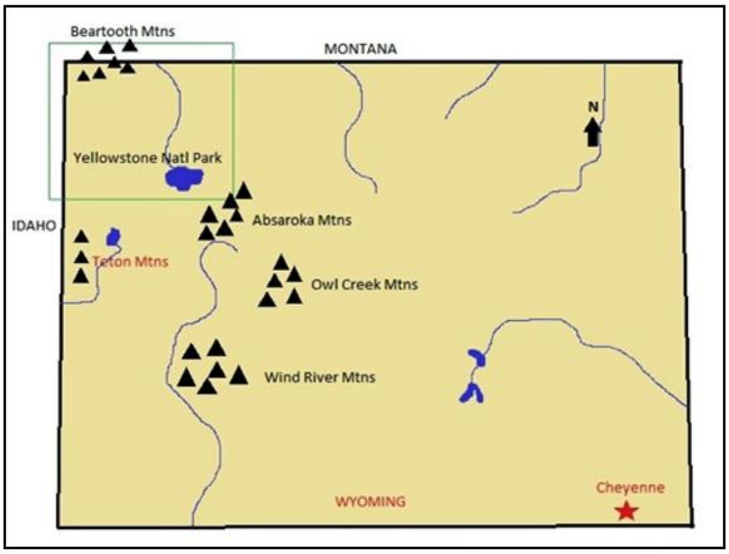

FIGURE 1. MAP OF WYOMING SHOWING THE LOCATION OF THE TETON RANGE AND SURROUNDING MOUNTAIN RANGES

Although the Teton Mountains are a young range, the rocks making up the core of the mountains are much older. The oldest exposed rock in the Teton Range is 2.8-2.7 Ga Archean gneiss composed mostly of quartz, feldspar, biotite, and hornblende (Reed and Zartman, 1973; Love et al., 2003). The backbone of the peaks is composed of the 2.5 Ga Mount Owen Monzonite, which is a granitic rock that contains 30-40% quartz, equal proportions (20-35%) of both potassium-rich and sodium or calcium-rich feldspar along with 5% or less biotite and traces of muscovite mica (Reed and Zartman, 1973). Irregular intrusions of pegmatite are found in granite exposures, which contain the same feldspars found in the Mount Owen Monzonite plus large flecks of muscovite and biotite mica and brown and red garnets (Love et al., 2003). The youngest Precambrian rock formation is an igneous diabase dike formation (Reed and Zartman, 1973). Sedimentary rocks aged 90-510 Ma are also present in the range and originated from the shallow sea that once covered Wyoming during the early Paleozoic (Craighead, 2006). The sedimentary stratigraphy around our study sites includes the GrosVentre Formation, the Bighorn Dolomite, and the Madison Limestone formations (Foster 1947; Love et al., 2003). Recent igneous formations include ash deposits from nearby volcanic eruptions. The Kilgore Tuff originated from volcanism 4.45 Ma associated with the Heise volcanic field in Idaho about 160 km southwest of Yellowstone National Park (Love et al., 2003). The Huckleberry Ridge Tuff formed 2.1 Ma by what is thought to be the largest known eruption from Yellowstone volcanoes (Fritz and Sears, 1993).

Since extension began to drive topographic uplift, the Teton Range experienced extensive glaciation; 12 glaciers still exist here today. Two primary glacial episodes in this region occurred during the Quaternary period. The Bull Lake glaciation occurred between 130-160ka, and the more recent Pinedale glaciation began more than 30ka and lasted until 14ka. Evidence of these glacial episodes include moraines, U-shaped valleys, ice-polished rock faces, large fans of outwash gravels, and kettle depressions left by melting blocks of buried ice. The extensive alpine glacial moraine deposits left behind from glacial retreat created natural dams, which formed Phelps, Jenny, Leigh, and Jackson Lakes (Pierce and Good,1992; Love et al., 2003).

III. METHODS

Fieldwork was completed over two field seasons in the summers of 2011 and 2012. All data were collected between July 25-August 13 of 2011 and July 27-August 5 of 2012. By collecting data at approximately the same time of year in both field seasons, error due to changes in seasonal weather were minimized as much as possible. All streams were east-flowing drainage systems of the Teton Range and included Paintbrush Canyon, Garnet Canyon, Death Canyon and Cascade Canyon (Fig. 2). These streams were chosen because they drained significant areas of the park (10-43 km²), allowed us to compare north-south variability, and were accessible by foot. The sampling locations span a north-south distance of 25 km. The average catchment lengths were approximately 7 km east to west. Paintbrush Canyon was the northernmost sampling area and Death Canyon was the southernmost sampling area

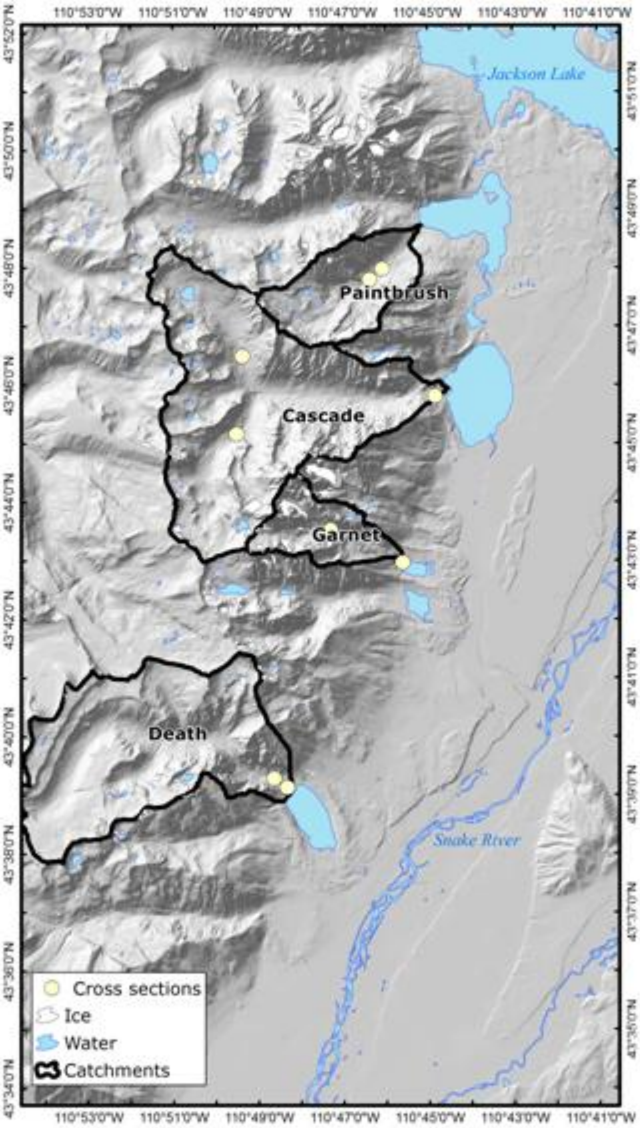

FIGURE 2. WATERSHEDS AND STREAM CROSS SECTION LOCATIONS

We collected field data to calculate discharge, which included measuring channel cross sections and observing width, depth and water velocity. We took flow velocity and depth measurements at equal intervals with a FP101 global flow probe across the width of the stream so that we collected at least 10-12 data points. We calculated cross section areas with the Reimann sum method of integration. We noted water surface slopes upstream and downstream of the cross section location. The slope of the water surface was found at each sampling location using a laser rangefinder.

We collected a surface bed load sample in each canyon by scooping sediment into a sample bag from within the active channel. Sediments finer than gravels and were sieved to calculate grain size distributions. At each sampling location, several gravel clasts were randomly selected and measured in the field with a granulometer. We collected 24 oz. water samples at each stream site to test for suspended sediments. A bottle held in a freely flowing section of the stream captured the water, and minimized organic materials floating in pools. In the laboratory, we filtered the water samples with 18.5 centimeter diameter P8-creped filter paper. We cleaned and dried bed load sediment samples with a vacuum and acetone. Then, we sieved sediments using three sieve sizes: 0.002 mm, 0.05 mm, and 2 mm.

We calculated stream competence based on basal shear stress and critical shear stress. Stream competence is defined as the largest size particle that a stream can carry. The basal shear stress, τ_b, is the force imparted on the streambed by moving water:

$$\tau_b = \Upsilon_w * R * S \tag{1}$$

where Υ_w is the specific weight of water (9800 N/m^3), R is the hydraulic radius (depth) of water, and S is the slope of the water surface. Critical shear stress, τ_c, is the theoretical force required to pick up a grain:

$$\tau_c = \Theta_{ec} (\Upsilon_s - \Upsilon_w) d \tag{2}$$

where Θ_{ec} is Shield's parameter for turbulent flow (0.044), Υ_s is the particle weight density (26,000 N/m^3), Υ_w is the specific weight of water (9800 N/m^3), and d is the diameter of the grain (m). We set the basal shear stress and the critical shear stress equal to each other to solve for the maximum grain diameter that a stream can carry. If the basal shear stress value is greater than the critical shear stress value, then entrainment of the sediment grains occurs in the stream channel. If the basal shear stress is less than the critical shear stress value, then deposition occurs.

Precipitation data for each watershed was obtained from the Parameter elevation Regressions on Independent Slopes Model (PRISM). This model uses point climate measurements from a network of weather stations to create a continuous grid of precipitation using a linear climate elevation regression function. This model showed a bulls-eye pattern with the highest precipitation values at topographic highs in the Teton Range and decreasing precipitation values away from the highest peaks (Foster et al., 2010).

IV. RESULTS

4.1 Discharge

We measured all stream cross-sections in the same locations during field work from July to August in 2011 and 2012 except on the north fork of Cascade Canyon. The cross section of the north fork was ~4 meters upstream in 2012 from where it was located in 2011 due to lack of accessibility, flow levels and vegetation growth. Many streams were wider or deeper during the high discharges in 2011 than in 2012 (Table 1).

Discharge values were compared over two years to evaluate how annual precipitation affected the measurements. The precipitation estimates and discharge measurements for each catchment are plotted together in Figure 3. Yearly precipitation data for the mountain range obtained from the weather warehouse database (weather-warehouse.com) was compared to model precipitation values based on elevation in Foster et al. (2010). From this comparison, we approximated the precipitation for each catchment per year. A qualitative comparison during our time in the field showed that there was more snow coverage in 2011 than in 2012.

TABLE 1
FIELD OBSERVATIONS, CALCULATED GRAIN SIZES AND DISCHARGE RESULTS

Stream	Sample ID[a]	Width (m)	Depth (m)	Slope	Basal shear stress	Calculated maximum d[b] (mm)	Observed maximum d[b] (mm)	Discharge (cms)
Cascade Confluence	AR11-23	17.42	0.37	0.071	258	-	-	12.26
Cascade Confluence	AR12-11	16.4	0.29	-	187	-	-	8
Cascade North	AR11-22	7.72	0.27	0.058	152.45	338	4096	3.93
Cascade North	AR12-10	8.99	0.19	0.022	41.36	111	1024	1.32
Cascade South	AR11-20	8.46	0.23	0.006	12.72	30	1024	2.23
Cascade South	AR12-9	8.48	0.13	0.008	10.23	25	256	0.71
Death Canyon	AR11-10	11.33	0.25	0.217	533.47	1453	90	8.78
Death Canyon	AR12-1	9.14	0.24	0.028	64.57	178	512	2.2
Upstream Garnet	AR11-18, AR11-19	7.9	0.14	0.01	14.53	31	64	1.45
Upstream Garnet	AR12-5	10.08	0.11	0.038	42.2	97	256	1.08
Downstream Garnet	AR11-13	9.8	0.19	0.024	43.68	119	90	2.12
Downstream Garnet	AR12-16	5.33	0.15	0.025	37.79	67	256	0.86
Paintbrush	AR11-15	6.27	0.25	0.091	223.92	497	90	4.72
Paintbrush	AR12-8	5.49	0.21	0.021	43.47	148	256	1.37

[a]AR11 samples collected in 2011. AR12 samples collected in 2012.
[b] d is grain size from equation 2.

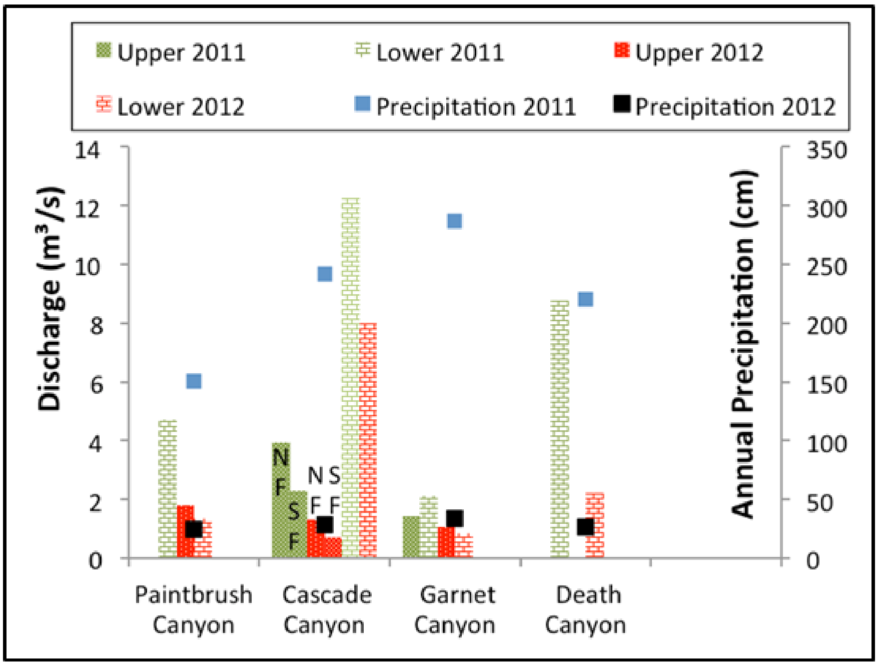

FIGURE 3. STREAM DISCHARGE COMPARED WITH ANNUAL PRECIPITATION. THE FOUR CANYONS STUDIED WERE PLOTTED FROM NORTH ON THE LEFT TO SOUTH ON THE RIGHT. NF REFERS TO NORTH FORK AND SF REFERS TO SOUTH FORK IN CASCADE CANYON

Elevation modeled precipitation across the range is low at the north end, increases to a maximum in the center between Garnet and Cascade Canyons, and then decreases again to the south. Stream discharges for 2011 were consistently higher than 2012, reflecting 35% higher precipitation in 2011 (Fig. 3). The highest precipitation values in both years were in Garnet Canyon. This canyon is most centrally located and also drains the highest elevations.

Although the canyons vary in size, we expected to see more discharge where precipitation was greater. In 2011, when the yearly precipitation was higher, the pattern of higher discharge with higher precipitation matched well, excluding Garnet Canyon (Fig. 3). In 2012, when the yearly precipitation was lower, the discharge was similar in most of the catchments.

We included a comparison of upstream and downstream discharges within each catchment. We expected that streams would gain water downstream as the area drained increased and more water entered the stream system. However, the data show there is not a constant relationship between discharge rates upstream to downstream. Most downstream sampling sites had higher discharge than upstream sites. Cascade Canyon followed the expectation that the downstream discharge would be greater. The trunk channel discharge was greater than the sum of the north and south fork channels combined. Additional smaller tributaries may enter the Cascade channel between the up and downstream cross-section locations. Paintbrush and Garnet Canyons were exceptions to the predictions of higher discharge downstream. Paintbrush and Garnet Canyons were smaller catchments than Death or Cascade Canyons. They have fewer tributaries contributing to discharge. With the smaller size, they may also be more easily influenced by fluctuations in discharge due to the time of day when sampling occurred due to snowmelt as temperatures fluctuated throughout the day.

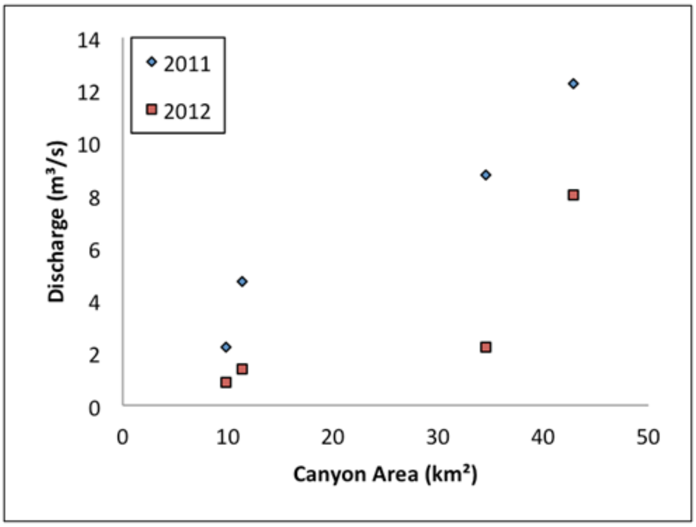

FIGURE 4. CANYON AREA COMPARED TO DISCHARGE FROM THE DOWNSTREAM SAMPLE SITES

If elevation controlled precipitation is not influencing the pattern of discharge across the range, we expected to see canyons with greater area supplying more water and resulting in greater discharge. This expected trend is very clear when comparing the downstream discharges in 2011 when there was greater precipitation across the range (Fig. 4). It still holds true in 2012, but the variability in observed discharge decreased. Garnet and Paintbrush Canyons are smaller than Death or Cascade Canyons; however, the 2012 discharges were similar to Death Canyon. Garnet Canyon discharge in 2012 was similar to the larger canyons when precipitation was lower, reflecting the importance of cooler high elevation temperatures delaying snowmelt and melting glacial ice contributions.

4.2 Sediment transport

While in the field, we mostly observed clear flowing water in the Teton streams. We observed low values of suspended sediment in our water samples and little to no clay or silt in the sieved sediments. We observed the most suspended solids in Garnet Canyon (Fig. 5). These observations may be related to the size and active processes in Garnet Canyon compared to the other canyons. Garnet Canyon has more melt water contribution from glacial melt than the other canyons in this study. Death Canyon showed the largest difference from year to year. The larger variation from year to year in Death Canyon could be caused by additional runoff sediment related to overland flow following a precipitation event.

There was little variation in the grain size distributions except for in Garnet Canyon (Fig. 6). The bulk weight percent of sieved sediments collected from all five canyons were fine to medium sand (0.075-2 mm).

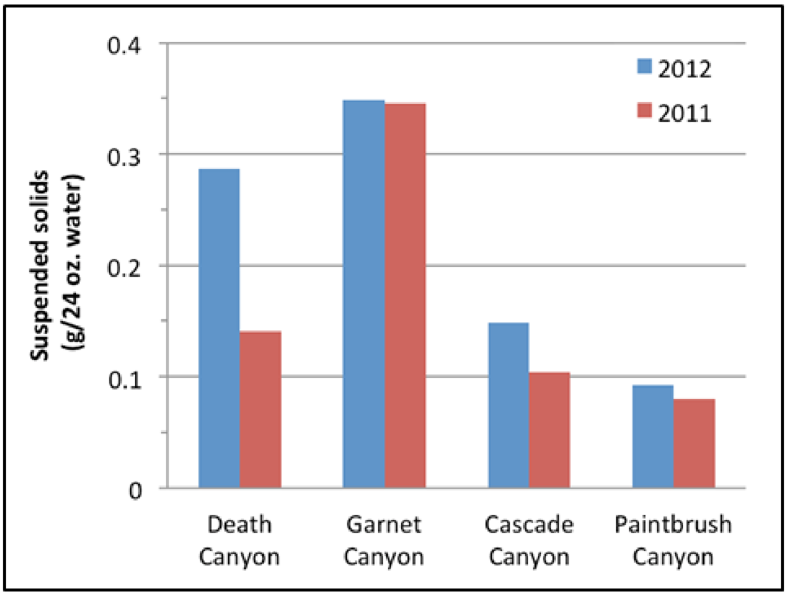

FIGURE 5. SUSPENDED SOLIDS OBSERVED IN EACH STREAM WATER SAMPLE

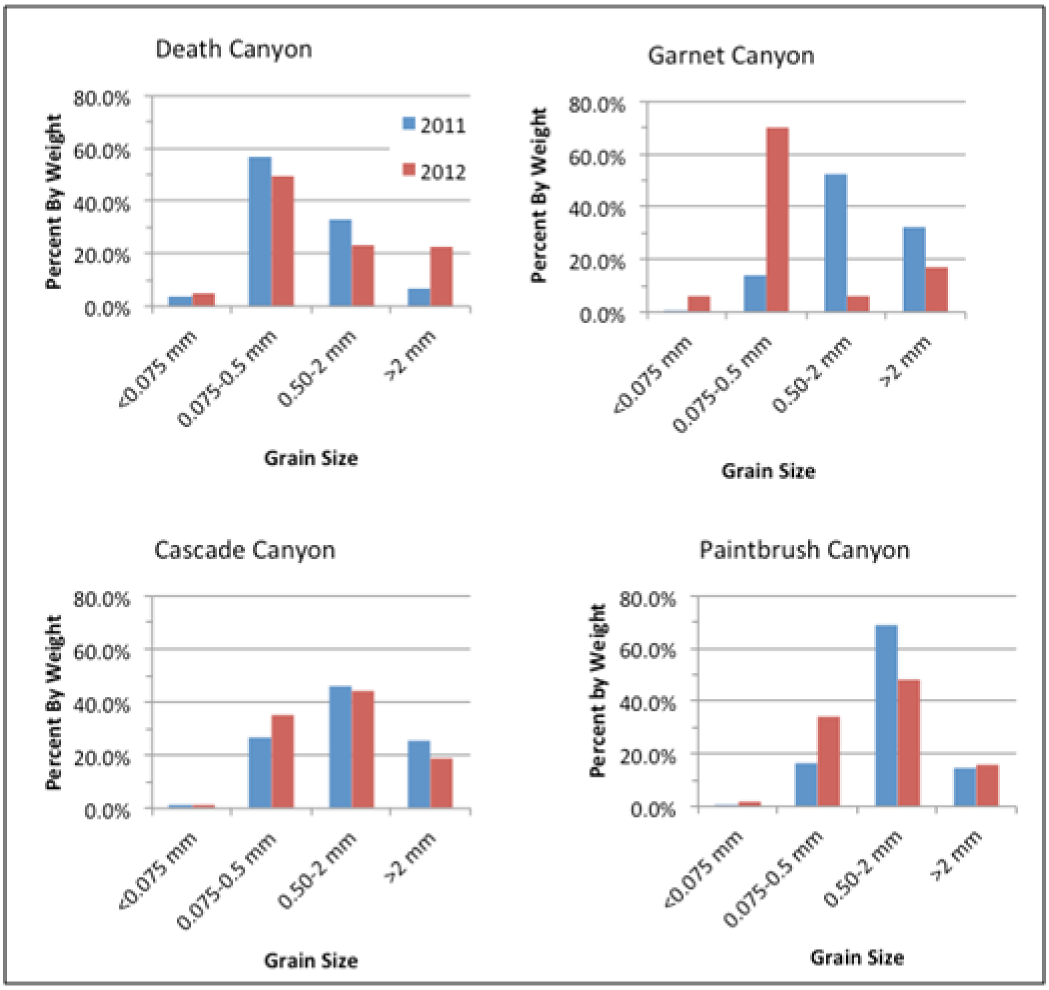

FIGURE 6. SIEVED SEDIMENT SAMPLES LESS THAN 2 MM

Paintbrush and Cascade Canyons, the more northerly canyons, contained more medium sand than the two canyons to the south.

Our field observations of coarse surface bedload measured with the granulometer showed a trend of larger clasts in larger watersheds in 2012, but that trend was less clear in 2011 (Table 1).In both years, the largest clasts were observed in the north fork of Cascade Canyon. Garnet Canyon contained the smallest clasts of all the stream locations sampled. Other than in Cascade Canyon, the larger grains were observed in 2012, the year with lower discharges and precipitation.

To investigate the effect of stream discharge on sediment transport, we predicted the maximum grain size transported under the observed conditions. The largest predicted clast size entrained by each stream was calculated when basal shear stress and critical shear stress were equal using basal shear stress in the deepest section of the stream. We compare our results in a log-log plot to assess similarities between our predicted and observed sediment measurements (Fig. 7; Barry et al., 2007). In general, when discharge was higher, we predicted larger clasts could be transported. In all but three of the cross sections, the maximum grain size we observed was larger than the predicted grain size. The exceptions were all from 2011 observations. One outlier was in Death Canyon, where although discharge was the second highest of the catchments, the sediment sizes were the smallest collected. This could be related to challenges collecting sediments from the fastest parts of the stream under high velocity conditions. The Paintbrush and downstream Garnet Canyon sites also resulted in higher calculated grain sizes than we observed in the field.

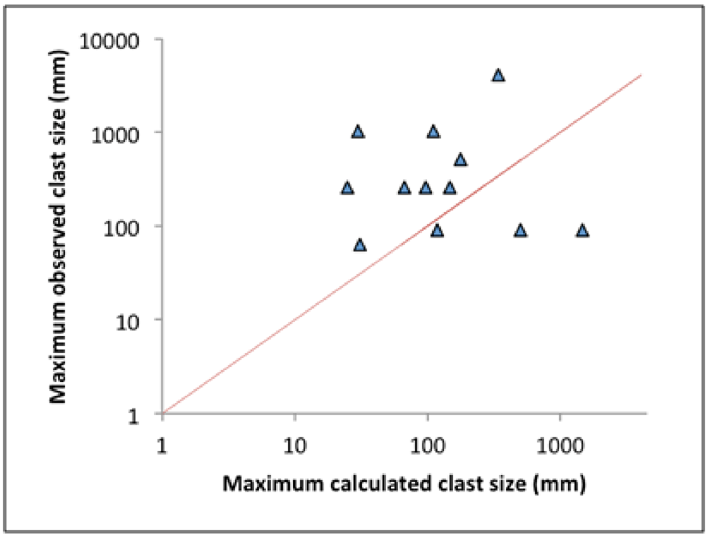

FIGURE 7. COMPARISON BETWEEN OBSERVED AND CALCULATED (PREDICTED) MAXIMUM GRAIN SIZES. THE LINE REPRESENTS WHERE THE SIZES WOULD BE EQUAL. ABOVE THE LINE, OBSERVED GRAINS ARE LARGER THAN CALCULATED GRAINS. BELOW THE LINE, CALCULATED GRAINS ARE LARGER THAN OBSERVED GRAINS

V. DISCUSSION

Annual precipitation and temperature conditions directly impact the volume of water flowing through the stream systems in the Teton Range. A forty-year study in the nearby Wind River Range concluded that the timing of snowmelt controls mountain stream discharge (Hall et al., 2012). Studies modeling how changing climate will affect snowmelt discharge predict that discharge will decrease, specifically during warmer seasons (Wang et al., 2016). In our study, we observed that higher precipitation and cooler spring temperatures in 2011 produced higher July stream discharges than in 2012. Although our results differ from the modeled results for long-term climate change, they agree with the expected variability associated with anomalous weather conditions in a given year (Hall et al., 2012).The implications are that more snowfall in the winter and spring may prolong the time within a given year when sediments can be transported efficiently and contribute to incision where bedrock is exposed, because, as previous studies have shown, high flow leads to greater channel erosion (Leopold and Maddock, 1953; Park, 1977).

Sediments in the Teton mountain streams are effectively transported by the flows we observed; therefore we also expect the streams are capable of incising bedrock. Our observed sediments were mostly larger than the predicted sediments that could be transported in the system, however, longer observation and sampling during flood conditions would probably prove these sediments capable of moving as well (Barry et al., 2007). Sklar and Dietrich (2001) explained that course sediments are more effective tools for erosion than fine sediments. Because we predicted that course grain sizes could be entrained by the observed stream flows, we reason that the sediments at our study sites were coarse enough to contribute to erosion. The 2011 discharges carried more suspended sediments; we observed more suspended sediments in our water samples and fewer sand-sized sediments deposited on the channel bed. Sampling at the same time in 2012 resulted in fewer suspended sediments and more sand deposition in the channel. The change from 2011 to 2012 was probably due to snow melt occurring earlier in the year in 2012.We surmise that incision was greater upstream than at our field sites because the beds we observed did not show exposed bedrock. Our downstream sample sites were near the stream mouth where, in general, more deposition was probable. The upstream cross sections likely fall in areas of coarse-grained deposition within a system of alternating coarse and fine channel deposits described by Wohl (1998). The grain size differences we observed from year to year may result from the different discharges, but are more likely related to potential uncertainties with our sampling technique. Comparisons between sediment sample observations in mountain streams show there can be much variability related to patchy sediment deposition (Bunte et al., 2012).

As expected, our observations show that larger catchments have greater potential for higher discharge, larger sediment transport, and therefore greater incision, but an interesting trend stands out that may be driven by relief within the range. The relationship between catchment size and discharge was stronger when the entire range experienced greater annual precipitation. When annual precipitation was lower, however, July discharge in Death Canyon was less distinct from the smaller Paintbrush and Garnet Canyons (Fig. 2). Garnet and Cascade Canyons border the highest peak (Grand Teton, elevation 4197 m) in the mountain range. As a result, areas within the catchments experience the greatest precipitation, coolest temperatures, and latest seasonal snowmelt. Death Canyon elevations are lower; therefore when annual precipitation is less, discharge, sediment transport, and incision are reduced in late summer.

Elevation also affects suspended sediment observations in the Teton Range. We observed more suspended sediments in Garnet Canyon than the other catchments, and those values were equally high both years. These results are possibly driven by two different explanations. One explanation could be related to relief within the catchment. Higher relief drives more efficient erosion and sediment transport (Walling, 1983). The Garnet Canyon stream channel has a steeper gradient than other streams in our study (Foster et al., 2010). Greater incision in upstream bedrock within Garnet Canyon may be possible due to the excess relief along with sufficient gravel tools in the bed load to abrade the stream channel. The second explanation for Garnet Canyon's higher suspended sediment load is related to the source of water within the channel. The Middle Teton Glacier in the north fork and a smaller glacier in the south fork of Garnet Canyon continuously feed some water to the stream channel in addition to snowmelt. Fine sediments trapped in glacial ice from erosion are released along with melting water (Hallet et al., 1996). Some of the suspended sediments we observed may be contributed by glacial melt water, however studies of lake sediments in Bradley Lake and Jenny Lake at the mouths of Garnet Canyon and Cascade Canyon respectively indicate that glaciers have not provided a significant contribution of sediments since retreat ~11 ka (Larsen et al., 2016).

VI. CONCLUSION

The results of our study begin to capture a record of stream discharge for channels draining the eastern flank of the Teton Range. Our findings also help us understand the potential for erosion in these mountain streams. We observed discharge and sediments during high flows in 2011 related to snowmelt, however they were not flood conditions. Late summer snowmelt directly influences stream discharge in mountain streams, which, in turn, allows for efficient sediment transport. Landscape features, including relief and elevation, may influence discharge and erosion potential more than catchment area in the Teton Mountains because snowmelt from high elevation can be prolonged during warmer years with lower precipitation.

This work begins an assessment of flow conditions in the Teton Range; however there remains much to be done to understand these mountain stream systems. More intensive monitoring throughout the year for several years would provide a more precise evaluation of the changes related to seasonal changes and individual precipitation events. Continuous monitoring or consistent monitoring at the same time of day will prevent discrepancies in data related to fluctuations in discharge caused by changes in daily temperatures. We also see interesting trends possibly related to elevation and catchment area, therefore a survey of many more streams in the range would help assess how the transport and erosion dynamics may be different related to size and relief. Lastly this study could be improved by more intensive or accessible sediment sampling strategies. In the faster 2011 flows, we observed smaller maximum grain sizes than in the slower 2012 flows. One possibility for this unexpected sediment size observation could be limitations to collecting in rapid and high flow (Wolman, 1954). When discharge is as high as we observed, it is possible that we were not able to collect in the fastest flowing area of the stream in 2011. If we were to use a sediment collector, we may be able to more actively characterize the sediments across the entire channel and capture all representative sediment size distributions.

Continued efforts to monitor discharge from mountain streams are important to understand the importance of snowmelt discharge in transporting sediments and supplying water resources as climate changes. Climate models indicate that snowpack in mountains in the western United States will change with a warming climate (Scalzitti et al., 2016). Those changes will thereby influence the timing of snowmelt discharge. Cooler temperatures tend to produce lower snowmelt discharges throughout the warm seasons as we saw in 2011, but overall warmer temperatures could result in continuous discharge as snowmelt throughout longer periods of the year, if any of the precipitation falls as snow at all (Molini et al., 2011). Snowmelt discharges from the Teton Mountains flow into the Snake River, which provides water to support agricultural irrigation and livestock (Clark et al., 1998). Variability in the timing of snowmelt discharge influences water availability for crops at different stages of the growing season, therefore it is important to understand the current dynamics of mountain streams because these systems are likely to change with changing climate in the future.

ACKNOWLEDGEMENTS

We thank Amber's thesis committee members Dr. Eric Peterson and Dr. Robert Nelson for their guidance, support, and reviews. We also acknowledge the University of Wyoming-National Park Service research station and Grand Teton National Park for assistance with access to field locations and accommodations. We thank Audrey Happel and Meredith Frisbee for field assistance. The Illinois State University College of Arts and Sciences New Faculty Initiative University Research Grant supported this research.

REFERENCES

[1] Barry, J.J., Buffington, J.M., King, J.G., 2004. A general power equation for predicting bed load transport rates in gravel bed rivers.Water Resources Research 40, W10401.

[2] Bunte, K., Potyondy, J.P., Swingle, K.W., Abt, S.R., 2012. Spatial Variability of Pool-Tail Fines in Mountain Gravel-Bed Stream Affects Grid-Count Results. Journal of the American Water Resources Association 48, 530-545.

[3] Clark, G.M., Maret, T.R., Rupert, M.G., Maupin, M.A., Low, W.H., Ott, D.S., 1998. Water Quality in the Upper Snake River Basin, Idaho and Wyoming, 1992-95, in: Survey, U.S.G. (Ed.), Denver, CO.

[4] Craddock, J.P., Kopania, A.A., Wiltschko, D.V., 1988. Interaction between the northern Idaho-Wyoming thrust belt and bounding basement blocks, central western Wyoming. Geological Society of America Memoir 171, 333-351.

[5] Craighead, C., 2006. Geology of Grand Teton National Park. Grand Teton Natural History Association, Moose, WY.

[6] Foster, D., Brocklehurst, S.H., Gawthorpe, R.L., 2010. Glacial-topographic interactions in the Teton Range, Wyoming. Journal of Geophysical Research-Earth Surface 115, F01007.

[7] Foster, H.L., 1947. Paleozoic and Mesozoic Stratigraphy of Northern GrosVentre Mountains and Mount Leidy Highlands, Teton County, Wyoming. AAPG Bulletin 31, 1537-1593.

[8] Fritz, W.J., Sears, J.W., 1993. Tectonics of the Yellowstone Hotspot Wake in Southwestern Montana. Geology 21, 427-430.

[9] Hall, D.K., Foster, J.L., DiGirolamo, N.E., Riggs, G.A., 2012. Snow cover, snowmelt timing and stream power in the Wind River Range, Wyoming. Geomorphology 137, 87-93.

[10] Hallet, B., Hunter, L., Bogen, J., 1996. Rates of erosion and sediment evacuation by glaciers: A review of field data and their implications. Global and Planetary Change 12, 213-235.

[11] Howard, A.D., 1998. Long Profile Development of Bedrock Channels: Interaction of Bedrock Weathering Mass Wasting, Bed Erosion, and Sediment Transport, American Geophysical Union Geophysical Monograph, pp. 297-319.

[12] Kirby, E.K., Whipple, K., 2001. Quantifying differential rock-uplift rates via stream profile analysis. Geology 29, 415-418.

[13] Larsen, D.J., Finkenbinder, M.S., Abbott, M.B., Ofstun, A.R., 2016. Deglaciation and postglacial environmental changes in the Teton Mountain Range recorded at Jenny Lake, Grand Teton National Park, WY. Quaternary Science Reviews 138, 62-75.

[14] Leopold, L.B., Maddock, T., 1953. They Hydraulic Geometry of Stream Channels and Some Physiographic Implications, in: Survey, U.S.G. (Ed.), Washington.

[15] Love, J.D., Reed, J.C., Jr., Pierce, K.L., 2003. Creation of the Teton Landscape. Grand Teton Natural History Association, Moose, Wyoming.

[16] Montgomery, D.R., Gran, K.B., 2001. Downstream variations in the width of bedrock channels. Water Resources Research 37, 1841-1846.

[17] Park, C., 1977. World-wide Variations in Hydraulic Geometry Exponents of Stream Channels: An Analysis and Some Observations. Journal of Hydrology 33, 133-146.

[18] Pickering White, B.J., Smith, R.B., Husen, S., Farrell, J.M., Wong, I., 2009. Seismicity and earthquake hazard analysis of the Teton-Yellowstone region, Wyoming. Journal of Volcanology and Geothermal Research 188, 277-296.

[19] Pierce, K.L., Good, J.D., 1992. Field guide to the Quaternary geology of Jackson Hole, Wyoming, Open-File Report - U. S. Geological Survey, Report 92-0504, p. 54.

[20] Reed, J.C., Jr., Zartman, R.E., 1973. Geochronology of Precambrian Rocks of the Teton Range, Wyoming. Geological Society of America Bulletin 84, 561-582.

[21] Roberts, S.V., Burbank, D.W., 1993. Uplift and thermal history of the Teton Range (northwestern Wyoming) defined by apatite fission-track dating. Earth and Planetary Science Letters 118, 295-309.

[22] Scalzitti, J., Strong, C., Kochanski, A., 2016. Climate change impact on the roles of temperature and precipitation in western U.S. snowpack variability. Geophysical Research Letters 43, 1-9.

[23] Sklar, L.S., Dietrich, W.E., 2001. Sediment and rock strength controls on river incision into bedrock. Geology 29, 1087-1090.

[24] Tomkin, J.H., Brandon, M.T., Pazzaglia, F.J., Barbour, J.R., Willett, S.D., 2003. Quanitative testing of bedrock incision models for the Clearwater River, NW Washington State. Journal of Geophysical Research 108.

[25] Tranel, L.M., Spotila, J.A., Binnie, S.A., Freeman, S.P.H.T., 2015. Quantifying variable erosion rates to understand the coupling of surface processes in the Teton Range, Wyoming. Geomorphology.

[26] Tranel, L.M., Spotila, J.A., Kowalewski, M., Waller, C.M., 2011. Spatial variation of erosion in a small, glaciated basin in the Teton Range, Wyoming, based on detrital apatite (U-Th)/He thermochronology. 23, 571-590.

[27] Walling, D.E., 1983. The Sediment Delivery Problem. Journal of Hydrology 65, 209-237.

[28] Wang, R., Kumar, M., Link, T.E., 2016. Potential trends in snowmelt-generated peak streamflows in a warming climate. Geophysical Research Letters 43.

[29] Wendel, J., 2015. World's Smallest Glaciers Risk Vanishing in a Warm Climate, EOS.

[30] Whipple, K.X., 2001. Fluvial Landscape Response Time: How Plausible Is Steady-State Denudation? American Journal of Science 301, 313-325.

[31] Whipple, K.X., Hancock, G.S., Anderson, R.S., 2000. River incision into bedrock: Mechanics and relative efficacy of plucking, abrasion, and cavitation. Geological Society of America Bulletin 112, 490-503.

[32] Wohl, E.E., 1998. Bedrock channel morphology in relation to erosional processes, in: Tinkler, K.J., Wohl, E.E. (Eds.), Rivers over rock: fluvial processes in bedrock channels. American Geophysical Union, Washington, DC, pp. 133-151.

[33] Wohl, E.E., Wilcox, A., 2005. Channel geometry of mountain streams in New Zealand. Journal of Hydrology 300, 252-266.

[34] Wolman, M.G., 1954. A method of sampling coarse river-bed material. Transactions - American Geophysical Union 35, 951-956.