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# SEDIMENT FLUXES IN A CHANGING CLIMATE: TAHOMA CREEK OVER DAILY TO CENTENNIAL TIME-SCALES

by

SCOTT WALLACE ANDERSON B.A., Lewis and Clark College, 2008

A thesis submitted to the Faculty of the Graduate School of the University of Colorado in partial fulfillment of the requirement for the degree of Master of Arts Department of Geography 2013 This thesis entitled: Sediment Fluxes in a Changing Climate: Tahoma Creek Over Daily to Centennial Timescales written by Scott Wallace Anderson has been approved for the Department of Geography

John Pitlick

Suzanne Anderson

Date

The final copy of this thesis has been examined by the signatories, and we Find that both the content and the form meet acceptable presentation standards Of scholarly work in the above mentioned discipline.

#### Abstract

Anderson, Scott Wallace (MA., Geography)

Thesis directed by Professor John Pitlick

Sediment fluxes in a changing climate: Tahoma Creek over daily to centennial time-scales

At Mt. Rainier, Washington, the intersection of climatic, glacial and volcanic processes conspire to create an extremely active landscape with the potential to respond dramatically to recent warming. In particular, perturbations to the frequency of the debris flows and floods that continually move mass from the mountain's flanks to downstream settings present a hazard to both the visitors and infrastructure of Mount Rainier National Park and downstream communities.

This work focuses on Tahoma Creek, a 40 km<sup>2</sup> basin on the southwest flank of Mt. Rainier. In the first chapter, we investigate fluvial transport processes occurring in the lower reaches of Tahoma Creek, below the influence of debris flows. Three aerial LiDAR datasets are used to create morphologic budgets that document channel change throughout the basin. We translate that change into estimates of bed load transport, and then present a method for interpolating a sediment rating curve for daily transport from this data. We confirm that our ability to predict sediment loads in steep streams is generally meager. Importantly, we show that, while equations generally over-predict transport at low to moderate flows, they significantly under-predict transport during very high flows, with large potential impacts on estimates for net sediment transport when considered over decadal to centennial scales. In the second chapter, we focus on the recent and historical sediment fluxes within Tahoma Creek in the context of geomorphic change. We first document channel change in response to recent debris flows, using LiDAR and dendrogeomorphic methods. We then use a suite of historic records to assess if any long-term trends in the morphology of Tahoma Creek are present that may reflect changes in regional climate. Finally, we construct a chronology of debris flows and floods within the basin since c. 1500 using tree core records from valley-floor conifers. We examine trends in debris flow frequency, and assess which climatic or glacial processes may control that frequency. We show that debris flows occur predominately during the onset of glacial retreat, though whether this reflects changes in sediment availability or changes in outburst flood frequency is unclear.

#### Acknowledgements

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# Using repeat LiDAR to measure sediment transport in a steep stream

Scott Anderson, John Pitlick and Paul Kennard

#### Abstract

Abundant sediment sources in paraglacial settings are initially mobilized and transported by steep mountain streams. As such, the transport rates in these streams set the pace and timing of downstream responses to changes in upstream sediment availability, making them an important control on hazards associated with these processes. However, our understanding of sediment transport in steep streams is limited by the inherent difficulty of measuring these fluxes. In this work, we demonstrate the potential of repeat topographic surveys to overcome this limitation, making use of aerial LiDAR surveys from the flanks of Mount Rainier in Washington State to quantify sediment fluxes during debris flows and fluvial transport. We focus on the  $38 \text{ km}^2$  Tahoma Creek watershed, with LiDAR surveys from 2002, 2008, and 2012. Making use of the fact that these surveys encompass all coarse sediment sources, we translate geomorphic change into total bed load transport volumes for the time steps between surveys. We then assume that the relationship between total daily sediment transport and daily mean discharge is of the form  $q_s = a(q-q_c)^b$ ; using our two observed total loads and estimates of discharge,

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we numerically solve for values of a and b, creating a bed load rating curve for Tahoma Creek. Comparisons with regional transport rates and deposition within a downstream reservoir indicate that our bed load transport estimates and derived rating curve are reasonable. This rating curve illustrates the impact of a large flood in 2006 - nearly 75% of the 2002-2012 load, and 30-40%of all predicted transport since 1943, occurred during this event. We use our estimates of bed load transport and derived rating curves to assess the performance of several bed load transport equations in these steep settings. The equations generally over-predict transport at low to moderate flows, but significantly under-predict transport rates during intense flooding. The use of a recently developed formula for critical shear stress improves agreement for lower flows, while shear-partitioning techniques designed to account for form-drag losses significantly under-predict transport at all flows. In quantifying the uncertainty in our measurements, we present an argument that cell-level survey errors have a minimal impact on our volumetric estimates of bed load transport, while potential survey mis-alignments provide the majority of our uncertainty. Propagating this uncertainty through to our rating curve, we show that our estimates of transport are most accurate for high flows, performing best in situations where traditional methods of bed load measurement generally struggle.

#### 1. Introduction

Steep mountain streams are energetic systems that transport significant volumes of sediment and are often the source of the majority of the coarse load within a basin. In areas subject to glaciation, the onset of glacial retreat further enhances this transport through the exposure of significant volumes of unconsolidated sediments (Church and Ryder, 1972; Hallet et al., 1996). Understanding the processes that mobilize and transport this sediment, collectively known as paraglacial processes, has become increasingly relevant as glaciers around the world continue to retreat (Haeberli et al., 1999; Paul et al., 2004; Barry, 2006). Specifically, it is these steep streams that link source areas to downstream systems, and so it is transport through these reaches that control the timing and pace of the downstream geomorphic response to changes in upstream sediment availability. To date, predicting these fluxes remains difficult - our current understanding of sediment transport in low-gradient rivers does not necessarily apply in these environments, where sediment transport equations often perform poorly. This poor performance has been variously explained as a failure to account for the increased form drag losses common in steeper streams (Chiari and Rickenmann, 2007; Nitsche et al., 2011), or the increasing critical shear stresses observed with increasing slopes (Mueller et al., 2005; Lamb et al., 2008). However, perhaps the most fundamental issue is the lack of observations of sediment transport in these steep streams on which to base empirical relations. Direct measurements of sediment transport, and particularly the geomorphically-significant

bedload component, are complicated by the coarse and energetic transport, coupled with the logistical difficulties of remote sites and episodic events.

In recent decades, the increased resolution and availability of topographic surveys has opened up the possibility of quantifying these fluxes by analysis of repeat measurements. This method, referred to as morphologic budgeting, has been shown to be effective for a variety of processes, including landslides (DeLong et al., 2012), debris flows (Schürch et al., 2011), and fluvial transport (Lane et al., 2003; Wheaton et al., 2010; Croke et al., 2012). However, there remains an open question as to what level of change can be detected in what environments, and so what processes can be meaningfully explored. Here, we present an analysis of repeat aerial LiDAR covering the active flanks of Mount Rainier, a volcano of the Cascade Range located in Washington State, USA. With these surveys, we quantify sediment fluxes over the entirety of a highly active, glacially-headed basin. Given that the basin has minimal obscuring vegetation or submerged topography, and undergoes geomorphic change that commonly exceeds even pessimistic estimates of survey accuracy, it is particularly well-suited to these morphologic budgeting techniques. The surveys encompass both the sources and primary depositional zones for debris flows in the basin, as well as the fluvial reaches below. We quantify the sediment fluxes through both these process domains, but focus our attention primarily on the lower fluvial reaches. Given that fluvial sediment transport is driven by stream discharge, a continuous and measurable quantity, it is much easier to connect the observed change with process here than in areas dominated by debris flow transport, as the number of debris flow events and associated triggering conditions are largely unknown.

The significant and potentially increasing volumes of sediment transported through this basin (and others in the region) represent a potential hazard to downstream communities. Indeed, this exact concern has prompted extensive research on Mt. Rainier and other Cascade volcanoes (Copeland, 2009; Czuba et al., 2012a; Lancaster et al., 2012). However, in this paper, we do not directly address these concerns, instead focusing on the bed load transport processes at work. The broader geomorphic significance of this transport, and an analysis of geomorphic change in the basin over decadal and centennial timescales, constitute the second chapter of this work.

#### 2. Study Area

The focus of this study is Tahoma Creek, a 38 km<sup>2</sup> basin draining the southwest flank of Mt. Rainier. Tahoma Creek is primarily sourced from South Tahoma Glacier, with minor contributions from a tongue of the Tahoma Glacier. Its first two kilometers flow through unconsolidated neo-glacial sediments, which provide much of the coarse sediment load for the river. The channel enters a confined reach between RKs 8 and 9, after which the valley widens significantly.<sup>1</sup> Debris flow deposition is focused within this wider

<sup>&</sup>lt;sup>1</sup>Locations with in the channel are denoted as the number of kilometers upstream from the Tahoma Creek bridge, following the valley centerline. The bridge then sits at river kilometer (RK) 0, the 2012 terminus at RK 13, and the confluence with the Nisqually river at RK -0.5. (figures 5, 6)



Figure 1: Map showing regional location of Mount Rainier, with inset showing the southern flank on the mountain. Tahoma Creek is the western-most of the indicated drainages. Letters indicate location of photos seen in figure 2

reach between RKs 4 and 8. Below this, fluvial processes dominate. In the lower reaches, the stream is primarily single-threaded with zones of mild braiding, flowing through an active channel 30 to 70 meters wide. This active channel is composed largely of bare gravel, hosting occasional alder stands. The channel is bordered by a 2 to 3 meter-high surface composed of fluvial overbank sediments and lahar deposits, supporting extensive conifer forests. Tahoma Creek has been notable for its high frequency of debris flows, with 27 recorded events over the past 40 years. The majority of these events occurred within two pulses, the first between 1967 and 1972, and the second



Figure 2: Photos of Tahoma Creek. Photo A is looking upstream towards the location where debris flow deposition generally begins. Valley-bottom forests that have been killed by recent debris flows are visible. Photo B shows a representative view of the downstream, fluvial reaches of Tahoma Creek. Locations of photos are indicated in figure 1

from 1986 to 1992. These debris flows have caused significant channel change in the upper basin and have forced Mount Rainier National Park (MRNP) to restrict access to sites along the western side of the Park. During the second of these pulses, USGS Cascade Volocanoes Observatory researchers Joe Walder and Carolyn Driedger were on site to document and characterize these events. They concluded that outburst floods from South Tahoma Glacier were the primary triggers for these debris flows, based on the frequency of events occurring during particularly warm, dry periods (Walder and Driedger, 1994b).

#### 3. Analysis of Repeat LiDAR

This analysis centers on three sets of aerial LiDAR, flown in 2002, 2008 and 2012. The two later datasets cover the entirety of Tahoma Creek from the glacial terminus to the confluence with the Nisqually River, while the 2002 dataset has coverage only above RK 1.3. After properly aligning these datasets, we perform the following steps:

- 1. Create 2002-2008 and 2008-2012 DEM's of difference (DoDs)
- 2. Sum volumetric change upstream of a point to provide an estimate bedload transport past that point, under the assumption that the LiDAR captures all coarse sediment sources and sediment may only exit the valley through down-stream motion
- 3. Use these estimates of total bedload transport, as well as flow records for the basin, to produce a simple sediment rating curve that allows us

to predict total daily bed load transport as a function of discharge

In our discussion of the potential error of these analyzes, we explore the relative impacts of cell-level uncertainty versus survey alignment uncertainty. We demonstrate that, in the case of aerial LiDAR, it is potential alignment errors dominate the uncertainty of our volumetric budget estimates, while cell-level uncertainties largely self-cancel in summation.

#### 3.1. Description of LiDAR Datasets

The LiDAR datasets were collected by three different operators, and so are the product of three unique methods of acquisition, processing, and quality assurance. The 2002 data was collected in early December by Terrapoint (now a subsidiary of GeoDigital) using a proprietary LiDAR system. Allreturn point density for the survey was 2.4  $pts/m^2$ , with an overall stated vertical accuracy of ~10 cm. This dataset extends from the glacial source down to RK 1.3, and so does not include the lower channel or the bridge. The 2008 LiDAR was collected by Watershed Sciences over several flights in mid-September, using a Leica ALS50 Phase II laser system. The mean ground-point density 0.73  $pts/m^2$ . This low value, the result of the dense forest cover and rough terrain, underestimates point densities for the bare gravel surfaces that constitute the majority of our analysis - over these areas, average point densities are closer to 2  $pts/m^2$ . Relative vertical accuracy was stated as 11 cm, while absolute vertical accuracy, as compared to 2240 RTK GPS survey points, was 3.7 cm with a mean value of 0.5 cm. The



Figure 3: A) Long profiles of change along the Westside road, used for vertical alignment testing. Distance is kilometers measured along along the roadway, with a sampling density of one point per meter. The 2008-2012 plot combines several discontinuous sections of the Westside Road as it leaves the valley around RK 7, and so extends significantly further than 2002-2008. B) Histograms of values shown in long profile plots above

2012 LiDAR was flown by the National Center for Airborne Laser Mapping (NCALM) between August 28th and September 1st, using an Optech Gemini Airborne Laser Terrain Mapper. The supplied report does not contain any quantitative information about point densities or relative or absolute accuracy. However, a plot of ground-point densities shows that most of the active channel areas contain 1-4  $pts/m^2$ , and accuracies are assumed to be similar to those of the 2008 survey.

#### 3.2. Co-referencing

The first step in measuring change through repeat surveys is to ensure that all the datasets are accurately co-referenced. As only the relative referencing is important for change analysis, we chose the 2008 LiDAR as our "ground-truth," and referenced both other datasets to it. All datasets were projected into NAD83 UTM coordinates, using the NAVD88 vertical datum and converted to orthometric elevations using GEOID03. Rasters were snapped and resampled to match the 2008 1-meter grid, producing exact cell-overlap between the datasets. Horizontal alignments were evaluated using a terrain-matching technique, similar to the method presented in DeLong et al. (2012). However, instead of using east-west and north-south transects to identify x and y offsets respectively, we calculate apparent change as a function of aspect over all areas presumed to be geomorphically stable - horizontal offsets create a sinusoidal pattern in a plot of mean change vs. aspect, with the peak/trough indicating the direction of the offset. We were able to detect and correct horizontal offsets of  $\sim 0.15$  m in both datasets. Vertical alignment was obtained by comparing the bare, level surfaces of several roads in the datasets (figure 3). After aligning both the 2002 and 2012 LiDAR to 2008 levels, we found an unresolved offset between the 2002 to 2012 road surfaces of 1.5 cm. Padding this value, we estimate our vertical alignment uncertainty to be 2.5 cm. Once aligned, sequential datasets were subtracted to create two DoDs, one representing change from December, 2002 to September, 2008, and one representing change from September, 2008 to September,

#### 2012.

#### 3.2.1. Impacts of Bare-earth Processing

The histograms of change along the road surfaces presented in figure 3 show two distinct distributions - change from 2002 to 2008 is normally distributed (skew = -0.15), while change from 2008 to 2012 shows a distinct positive skew (skew = 1.32). This positive skew reflects differences in how aggressively vegetation was stripped out from the original all-return point clouds to produce a bare-earth surface. The 2002 and 2008 datasets represents a relatively aggressive approach, resulting in smoother, but less dense, products, while the 2012 survey represents a less aggressive approach, in which more points are retained but there is an increased probability that some of these points are mis-classified. In a sense, the 2012 dataset uses a wider window to define ground points than the 2002 or 2008 datasets. Since points can not reflect below the true ground surface, the opening of this window does not increase the number of low points; only the number of high points increases. This asymmetry between the number of retained high and low points is the ultimate cause of the skew seen in change along the road surface, which is commonly obscured by vegetation. The relative lack of such vegetation along the fluvial channel limits how much of an impact this has on our final results, but such differences are likely common when surveys are performed by different operators.

#### 3.3. Assessing the Role of Survey Uncertainty

The topographic surveys we use to quantify change are inherently subject to some level of uncertainty. The majority of this uncertainty can be divided into two sources - cell-level uncertainty and global alignment uncertainty. Cell-level uncertainty represents the confidence that a given cell's elevation corresponds to the true elevation (or that the change seen in a DoD cell is the true change), while global alignment uncertainty represents the confidence that two datasets have been correctly aligned. Here, global alignment errors are taken to be only in the vertical direction, as these have the most direct impact on change analysis. We denote these uncertainties as normal distributions with standard deviations  $\sigma_c$  and  $\sigma_g$  respectively. Of the two, cell-level errors have received a majority of the attention in past works, with numerous attempts to quantify  $\sigma_c$  and how that value may vary across a landscape (Wheaton et al., 2010; Milan et al., 2011). These estimates often constitute a basis for defining a threshold of detection - changes smaller than some value are considered indistinguishable from noise, and so excluded from the analysis (Croke et al., 2012, e.g.,). However, in cases where the aerial extent of the surveys is large, and net volumetric change is the quantity of interest, we argue that it is global alignment that dominates the total error, and that the cell-level errors are relatively inconsequential. Most simply, this is because cell-level errors, assumed to be normally distributed around zero, should be somewhat self-canceling in summation, while alignment errors all operate in the same direction, and so increase linearly with area. We provide calculations below that quantify the relative impacts of these two sources of uncertainty to more clearly demonstrate this. In these calculations, we directly discuss the uncertainty in a derived DoD, noting that the cell-level uncertainty of the two parent surveys,  $\sigma_a$  and  $\sigma_b$ , combine such that  $\sigma_c = \sqrt{\sigma_a^2 + \sigma_b^2}$ 

### 3.4. Comparing Cell-level and Alignment Uncertainty

We begin with the purely statistical statement that the mean value of n observations pulled from a normal distribution approximates the true mean with an uncertainty defined by the standard error of the mean,

$$\frac{\sigma}{\sqrt{n}}$$

By extension, the *sum* of n observations pulled from a normal distribution approximates the expected sum,  $n\mu$ , with an uncertainty

$$\frac{\sigma}{\sqrt{n}}n=\sigma\sqrt{n},$$

a value we refer to here as the standard error of the summation. Summing normally-distributed random values is exactly what occurs when cell-level errors are summed in volumetric estimates of change. In this situation, n should most properly be the number of independent xyz points. However, we choose here to use the number of cells in a raster - in the case of 1m DEMs, this simply becomes the area in square meters. Given a point density of ~1 pt/m<sup>2</sup>, the two values are equivalent. The relevant standard deviation is that of the cell-level uncertainty,  $\sigma_c$ . Using these values, the standard error of the summation then becomes a direct estimate of the volumetric uncertainty associated with cell-level errors. If we then divide this number by the area of analysis, we are are left with a length-scale that defines the effective mean vertical uncertainty of the summed cell-level errors. More simply, it defines the global alignment uncertainty that would produce an equivalent volumetric uncertainty over the area of analysis. With this value, which is simply the standard error of the mean we started with, we are then able to directly compare the two sources of error, as seen in figure 4.

Note that it is possible to logically extend this analysis to rasters of any cell size, as long as the assumption of  $\sim 1$  pt/cell is reasonable - in the case of a DEM with cell size L,

$$\frac{\sigma_c}{\sqrt{n}}nL^2 = \sigma_c\sqrt{n}L^2$$

provides a measure of the volumetric uncertainty for the area of analysis. The standard error of the mean, with n taken to be the number of cells, always provides the global alignment equivalent regardless of cell size.

#### 3.5. Application to this Study

To estimate  $\sigma_c$  for our DoDs, we used the standard deviation of several thousand observations of change along the West Side Road as seen in figure 3 - for both time periods, this was 0.08 m. For an area of 10,000 m<sup>2</sup>, cell-level



Figure 4: Standard error of the mean plotted for given areas, using indicated values of  $\sigma_c$ .

uncertainty of this magnitude contributes the equivalent of  $\pm 1$  mm of global alignment uncertainty, or  $\pm 10$  m<sup>3</sup> of total volumetric uncertainty; for an area of  $1.5 \times 10^6$  m<sup>2</sup>, it contribute the equivalent of  $\pm 0.1$  mm of global alignment uncertainty, or  $\pm 200$  m<sup>3</sup>. Given that we estimate our global alignment uncertainty at 2.5 cm, this additional uncertainty is essentially negligible. Further, 200 m<sup>3</sup> represents 0.01% to 0.2% of the net volumetric fluxes recorded in Tahoma Creek between '02-08 and '08-'12, respectively. For these reasons, we elect to quantify our uncertainty solely as a function of global alignment errors, and ignore cell-level errors. We also forgo applying any threshold of detection, as the logic above suggests that it provides no obvious benefit in our setting.

#### 3.6. Estimating Bed Load from DoDs

Based on both field observations and LiDAR analysis, the vegetated valley walls contributed no significant sediment to the channel over the 2002-2012 time period. As such, we restrict our analysis of geomorphic change to the active fluvial channel and immediately adjacent terraces that showed reworking consistent with fluvial action. This area was subdivided into 100m sections as measured along the valley centerline, and volumetric change calculated for each such section. The downstream trends in aggradation and incision, expressed as the mean vertical change of a section, are shown as the blue lines in figure 7.

Given that these surveys encompass the entirety of Tahoma Creek, and

that sediment may only leave the basin via downstream transport, we interpret the net volumetric change upstream of a point as the volumetric flux of sediment past that point, assuming that:

- 1. The source and sink bulk densities are the same
- 2. The LiDAR datasets cover all sources and sinks of coarse sediment
- 3. There is a clear distinction between areas of ice mass loss and sediment mobilization

Given that the source material is unconsolidated glacial debris, which has a bulk density similar to that of stream gravel, the first assumption is reasonable. The second assumption is met with the exception of sub-glacially sourced sediment - however, prior research on Mt. Rainier's glaciers has noted that sediment loads at the glacial termini are composed largely of fine material traveling in suspension, and that most of the coarse load is derived from entrainment of downstream sediments (Fahnestock, 1963; Mills, 1979). The last assumption is the most difficult to assess; for the 2002-2008 time period, there was  $\sim 300$  m of glacial retreat - within this area of retreat, it is not possible to differentiate between ice mass loss and the mobilization of sediment, and so was excluded from the analysis. This alone would mke the estimate of the load mobilized from this upper source area a lower limit. Conversely, photographs taken in 2012 show stagnant ice in the lower neoglacial sediments around valley km 11, and it is clear that some percent of the apparent incision in this area was into ice, inflating our transport estimates. While it is not possible to directly quantify the volumes of ice that

may have be eroded, previous work in the basin gives some guidance. Scott et al. (1995), examining debris flows in 1986-87, noted ice making up to 10% of the total depositional volume. Walder and Driedger (1994a) noted that the debris flows in the early '90s, while clearly incising channels into stagnant ice in the source zone, contained no visible ice or frozen ground in their deposits. While the depositional extent of ice is a minimum estimate of the total eroded volume, it provides some assurance that this volume is not unreasonably large. We crudely estimate that these errors represent no more than 25% of the total loads observed, and note that these two sources of error at least work in opposite directions.

Finally, we take these transport rates to be bed load transport rates, as opposed to total loads, an extension of the assumption that most of the suspended load is sub-glacially sourced and quickly exits the system with minimal local geomorphic impact. We also assume that the morainal sediments that constitute the basin's source area are primarily composed of coarse material similar in nature to Tahoma Creek's bed material. An analysis of sediments in moraines around Mt. Rainier showed that material finer than 2 mm made up less than 35% of the deposits,<sup>2</sup> with D50 values ranging between 32 mm and 128 mm (Mills, 1978). No samples were taken from moraines of either the South Tahoma or Tahoma glaciers, but moraines around the

 $<sup>^{2}</sup>$ This percentage is based on sediment distributions truncated above 102 mm. Inclusion of coarsest fraction would lower this percentage, but such data is not available within the referenced paper

mountain were quite similar, and so provide a reasonable approximation for our basin. In comparison, subsurface sediment samples taken along Tahoma Creek during the summer of 2012 were composed of 18% material finer than 2 mm, while D50s were between 55 mm and 90 mm, with a mean of 71 mm. These numbers show that the sediments stored in the upper source areas are not significantly different that the materials composing the bed within the lower channel. As it is changes in these surfaces that we ultimately measure, our methods should capture the majority of coarse sediment transfers occurring within the basin. Our estimates of bed load fluxes are presented as the black lines in figure 7. The plausible uncertainty of these values is presented as the transport volumes calculated after applying  $\pm 2.5$  cm and  $\pm 5$  cm of vertical offset to the two DoDs, shown as the grey bounding polygons.

#### 3.7. Description of Change, 2002-2008

The 2002-2008 time period is dominated by the extensive work of an extremely large flood in November of 2006. In the upper basin, debris flows mobilized over 2.3 million  $m^3$  of unconsolidated glacial sediments (figure 6). Of this, one million  $m^3$  was deposited between RKs 4 and 9, while the remaining 1.3 million  $m^3$ , along with 47,000  $m^3$  entrained in the lower valley, was transported beyond the extents of the 2002 survey (figure 7). Most of this sediment likely continued downstream to enter the Nisqually River. Below RK 4, the river generally incised on the order of 0.5 m, with the exception of a broad fan of deposition occurring around RK 3.3 (figure 6, 7). This depo-



Figure 5: Vertical change in the upper basin of Tahoma Creek. Less transparent coloring of change is used over the area of analysis.

sition follows a marked decrease in channel slope, and field evidence suggests this area is a common depositional zone during large transport events. Between RKs 3.5 and 8, repeated debris flows and the intensity of the 2006 flood cause the main thread of Tahoma Creek to move significantly over this period, ultimately abandoning it's course along the eastern margin of valley between RKs 3.5 and 4.5 to flow through a previously forested area to the west.



Figure 6: Vertical change in the lower basin of Tahoma Creek.



Figure 7: Mean vertical change and estimated sediment transport as derived from LiDAR differencing. Light and dark grey shading represents transport estimates for  $\pm 5$ cm and  $\pm 2.5$ cm of offset respectively

#### 3.8. Description of Change, 2008-2012

The later time period showed significantly less transport at all points in the basin. The upper reaches mobilized 250,000 m<sup>3</sup> of material, predominately sourced in sediments mantling a steep bedrock step just below the 2012 terminus (figure 5). Lateral moraines showed significant downslope motion of material, but the majority was deposited along the base of the slopes, with only a modest volume being mobilized into the lower system. Deposition, probably the result of small debris flows, began at RK 8.5, with 150,000 m<sup>3</sup> of material being deposited between this point and RK 7. Below this,
fluvial processes transported a consistent  $100,000 \text{ m}^3$ . While the channel in this lower area has been reworked, it remained essentially vertically stable over most of its length. The only major exception to this occurred below the Tahoma Creek Bridge, where 25,000 m<sup>3</sup> of material was deposited as the river widens at its confluence with the Nisqually River.

The relative volumes of transport for the two time periods are similar, with roughly half of the material mobilized from the upper basin continuing downstream to exit the system, leaving the other half deposited primarily between RKs 6 and 10. While the lower channel throughput large volumes of sediment and does show localized vertical change, there does not appear to be any broad trend of aggradation or incision over the period of record.

#### 4. A LiDAR-Derived Sediment Rating Curve

The total loads observed in our DoDs are the result of the integrated forces driving transport. For fluvial transport in the lower valley, this driving force is the stream discharge. Taking advantage of several gages operated in the Nisqually Basin, including a limited period of record for Tahoma Creek itself, it is possible to estimate this discharge over the past 50 years, encompassing the entire period between 2002 and 2012. With two unique total loads, and our estimate of the hydrology driving that transport, it then becomes possible to back out a simple two-parameter sediment transport rating curve. We choose to use a rating curve of the the form  $q_s = a(q - q_c)^b$ , a common form for simple bedload formulas (Meyer-Peter and Müller, 1948; Rickenmann, 2001). In this formula,  $q_s$  is the total daily load (kg/day), q is the daily mean flow (cms), and  $q_c$  is the critical discharge below which it is presumed no transport occurs. Setting the critical flow,  $q_c$ , to a reasonable value, it is then possible to numerically solve for the unique values of a and b that satisfy the paired equations

$$\int_{t_1}^{t_2} a(q-q_c)^b = TL_{02-08}$$
$$\int_{t_2}^{t_3} a(q-q_c)^b = TL_{08-12}$$

where TL indicates the total load over the noted period. The derived values of a and b then define the functional relationship between discharge and bed load transport in Tahoma Creek at a daily scale.

Inherent in this method is the assumption that the relationship between discharge and sediment transport has been stationary over the period of observation. Indeed, such assumptions underlie most observationally-based rating curves, where scatter over orders of magnitude is usually fit with a single curve taken to represent a mean state (e.g., Recking, 2013, fig. 2). In this case, as with most other cases, this assumption is essentially impossible to verify. Significant changes in the bed material or local bed topography represent additional potential sources of error within this analysis.

Below, we describe the methodology we used to estimate discharge near the Tahoma Creek bridge. This same discharge was used when calculating a and b values for all points along the basin. While this will produce overestimates of the flows in the upper reaches, the drainage area at RK 3 still is 85% of that at the bridge, and drops to only 70% by RK 4.5.<sup>3</sup> Above this point, the flow is split between the main thread of Tahoma Creek and Fish Creek, and so our discharge estimates are clearly in error. While we present the results above this confluence, their physical meaning is somewhat muddied, both because the discharge is likely unrealistic, and more importantly, because transport in these reaches occurs predominately in debris flows, and so is not driven directly by stream discharge.

#### 4.1. Streamflow

Direct measurements of discharge in Tahoma Creek were obtained from a gage at the bridge recording stage every 15 minutes from June 14th to August 24th, 2012. Discharge measurements, taken roughly once a week, were used to construct a rating curve for the site, although the highly mobile bed made this a non-trivial effort. Two gages on the Nisqually River provided more consistent and extensive records - one located at Longmire, operated by the NPS since 2009, and one located 27 kilometers downstream at National, operated by the USGS since 1943. The Longmire gage was washed out during a December 2010 flood, and was not replaced until June 2011, creating a

<sup>&</sup>lt;sup>3</sup>The exponent in our equation, which is more physically meaningful than the coefficient, is a function of the relative hydrology between the two time periods only, and not the absolute discharge. Thus, applying same simple area-scaling to discharges over both time periods does not change this value

significant gap. Further, the rating curve developed at Longmire for the period before this gap was not well defined for high flows. For this early period, we believe that low flow values are reasonably accurate, but consider peak flow data unreliable and so exclude it from analysis.

Flows measured at Longmire and National did not scale simply, reflecting the difference between the glacially-dominated flows at Longmire and the mixed glacial-lowland basin at National. To develop a relationship between these two regimes, flow was first separated into base- and event-flow components using a local-minimum method (Pettyjohn and Henning, 1978). These two elements were then analyzed independently. Monthly-mean base flow ratios (taken as Longmire/National) reflected the influence of summer melt from glacial ice and snow, with Longmire having a relatively high base flow during summer and early fall, and a relatively low base flow during the winter (figure 8a). Event flow ratios showed a bimodal distribution, with low values during the winter months, when much of the high-elevation precipitation falls as snow, and higher ratios during the melt and snow-free periods (figure 8b). Using these two relationships for base- and event-flows, we reconstructed discharge at Longmire for the entire period of record of the National gage. The results of this process, shown in 8c, provide improved agreement for flows during the melt season when compared to simple area-scaling. Given the similarities between the Nisqually at Longmire and Tahoma Creek at the bridge, we assume that the two flows are highly correlated. The Longmire-National relationship was then rescaled to match the drainage area of Tahoma Creek,



Figure 8: Various components of the synthetic hydrograph process, including A) unit baseflow (cms/km<sup>2</sup>) ratio for Longmire/National, showing all individual months and means for each calendar month, B) event flow (cms/km<sup>2</sup>) ratio for Longmire/National, shown in relation to SWE as measured at the Paradise SnoTel station, and a comparison of the derived synthetic hydrograph with the observed flow for C) Longmire, and D) Tahoma Creek. In sub-figure C, the results of simply scaling National discharge to reflect the drainage area of Longmire are also shown to highlight the improved performance of the synthetic method during the melt-season

and the resulting synthetic flow records compared to the measured discharge for the summer of 2012. While the flows showed good relative agreement, the synthetic discharge was consistently higher than the measured discharge by  $\sim 50\%$ . A simple scalar was included in the relationship to correct this offset, producing the final fit seen in figure 8d. To determine if our estimates for peak flows were reasonable, we compared our synthetic estimate of the 2006 flood to photos taking at the Tahoma Creek Bridge during and after this event. Using the bridge's width of 20 m and assuming the flow velocity to be 2-3 m/s, our predicted flow of 68 cms gives a depth of 1-1.7 m, in good agreement with the depths inferred from the photos.

#### 4.2. Critical Discharge

Since two time periods of analysis only give us the two total load equations shown above, we are only able to solve for two free parameters in our rating curve. As there are three such parameters (a, b and  $q_c$ ), we must choose one to fix before solving for the remaining two. We choose to solve for a and b using two plausible values of  $q_c - 0$  cms, representing un-thresholded transport described by a pure power-law, and a field-derived estimate of 5.5 cms. This value was based on the numerous shifts observed in stage-discharge relationship developed over the summer of 2012, which occurred after flows at or above this discharge. These shifts represent the restructuring of the bed, and so are a rough guide to mobility. This flow is equivalent to a critical shear stress of 0.07-0.09, in good agreement with several recently developed formulas for critical shear stress (figure 14; Mueller et al. (2005), Lamb et al. (2008)).

This single value belies the larger complexities inherent in bed mobility processes. At the onset of melt, several flows at or above this critical discharge caused no apparent bed restructuring, nor was any bed load transport detected during an early-season measurement taken at 7 cms. Only after a 7.8 cms flow caused initial bed mobility were lower flows observed to cause rating curve shifts. These results reinforce the pervasive opinion that the factors controlling thresholds of motion in gravel-bedded rivers are significantly more complex than most current models account for (Turowski et al., 2011). The observed increase in channel mobility also coincided with the onset of significant diurnal fluctuations in discharge, indicating the onset of the melt season. This melt flushes fines from the upper basin, and produced a remarkably linear decrease in channel roughness as this material filled interstitial grain space. This influx of fine material, and associated decreases in roughness, may also have contributed to the increased mobility of the bed.

# 4.3. Results and Uncertainty

With our synthetic discharge and two chosen values of  $q_c$ , we numerically solved for values of a and b in the equations above, producing sediment rating curves for any location with coverage in both DoDs. The resulting values of a and b for the un-thresholded and thresholded forms of the equation are seen



Figure 9: rating curve coefficients and exponents as derived from LiDAR analysis using the two indicated forms. Light and dark shading represent one and two standard deviations of uncertainty, as calculated from Monte Carlo analysis

in figure 9, with values calculated every 100m from RK 1.3 up to the glacial terminus. Exponents and coefficients for the un-thresholded form are around 3.2-3.4 and  $2-7\times10^3$  respectively, and 1.5-1.8 and  $1-3\times10^6$  for the thresholded forms. Both exponents and coefficients for both forms are consistent in the downstream direction, with identical spatial trends. Exponents on the higher end of the observed range coupled with coefficients on the lower end of the observed range vary little between RKs 1.3 and 7.5, at which point exponents decrease and coefficients increase smoothly between RKs 7.5 and 8.3. Above

this, values again vary little.

To estimate the uncertainty of these results, we performed a 500-run Monte Carlo analysis. For each run, two random vertical offsets were generated from a normal distribution ( $\mu = 0, \sigma = 0.025$ ) and added independently to the two DoDs, after which new total loads were estimated. The hydrology for both time periods was multiplied by a single random normally-distributed value ( $\mu = 0, \sigma = 0.2$ ). These quantities were then used to calculate new values of a and b, and the distribution of these values were taken to represent our uncertainty. As this only accounts for uncertainties in LiDAR alignment and our synthetic hydrology, and not the interpretation errors mentioned in section 3.6, it is a low estimate of the total uncertainty. 1- $\sigma$  and 2- $\sigma$  values are shown in figure 9 as dark and light shaded areas respectively. This uncertainty increases downstream as a simple function of contributing area, the result of quantifying our uncertainty largely as a function of imperfect survey alignments.

At present, we do not have any quantitative data to suggest which of the two forms of our rating curve better matches physical processes in Tahoma Creek. By design, both rating curves exactly predict the observed total loads between 2002-2008 and 2008-2012. Both give similar results when used to estimate transport loads over the period of our synthetic hydrology - 57,000 m<sup>3</sup>/yr for the thresholded form, compared to 43,000 m<sup>3</sup>/yr for the un-thresholded form. Both equations predict similar transport volumes for the 2006 event. The thresholded form does eliminate the need for a single

power-law to explain transport across flows ranging from 1 to 68 cms, and so feels more physically plausible. However, the use of a threshold adds curvature to the discharge-transport relationship in log-log space, as can be seen in figure 13. This curvature complicates comparisons between our rating curve and other estimates of transport expressed as power-laws, as well as to modern transport equations which generally do not contain similar thresholds. As such, we elect to present total loads as calculated using our thresholded rating curve, but use the un-thresholded form as a more intuitive comparison to rating curves derived from both bed load measurements and transport equations.

# 4.4. 1943-2012 Predicted Transport

The derived rating curves, in combination with our synthetic hydrology, allow us to estimate coarse sediment fluxes from Tahoma Creek between 1943 and present (figure 10). Year-to-year variations are significant, including some years where flows never exceeded the critical discharge. Looking at the cumulative transport over time, the period from 1943 until the late '70s shows a relatively consistent transport rate. Between the late '70s and early '90s, transport rates are generally low, with only two years over transport rates near the mean of the period. This quiescent period ends in 1996, with the second largest transport event of record. From 1996 to 2012, cumulative transport rapidly increases, most significantly due to the 2006 flood. This single event represents 30% of all transport predicted since 1943. While this



Figure 10: Annual coarse material transport at RK 1.3 since 1943, as derived from thresholded rating curve. Cumulative transport is shown against right axis. Note that 2006 transport rates have been truncated.

later period encompasses several extreme events, transport rates outside of these events are similar to what was seen before the late '70s.

# 4.5. Assessment of Results

To assess whether our estimates of bedload transport and the derived rating curves are reasonable, we compare our results to prior estimates of sediment transport, focusing on research around Mt. Rainier.

Period	Volumetric Deposition (m3/yr)	Bed load Volume (m3/yr)	Tahoma Creek Transport (m3/yr)	Percent of Total
1956-1985	430,000	86,000	25,000	29%
1985-2011	770,000	154,000	73,000	47%
1956-2011	580,000	116,000	47,000	41%

Table 1: Comparison between Alder Lake deposition and predicted Tahoma Creek transport. Bedload is take to be 20% of the total depositional volume

#### 4.5.1. Additional LiDAR-Derived Bedload Estimates

The most direct comparison of our results comes from LiDAR-derived bed load transport estimates from the Kautz, Carbon and Nisqually basins, all pro-glacial catchments around Mt. Rainier similar in size to Tahoma Creek. As all of these basins are covered by 2008 and 2012 aerial LiDAR survyes, we used the same methodology outlined for Tahoma Creek to produce estimates of the coarse sediment transport over this period. At the downstream extent of the surveys, the Carbon River transported  $1.2 \times 10^5$  m<sup>3</sup>, while the Nisqually River transported  $0.8 \times 10^5$  m<sup>3</sup>. The uncertainty in these values is comparable to that seen for Tahoma Creek transport of this same period (figure 7), but they are very similar to the best fit values of  $\sim 1 \times 10^5$  m<sup>3</sup> seen in Tahoma Creek. Kautz Creek, which is almost entirely armored with coarse sediment from 1947 debris flows, showed no detectable transport.

#### 4.5.2. Alder Lake Sedimentation

Alder Lake, a reservoir on the Nisqually River in operation since 1945, provides a means for estimating total loads for the whole of the upper Nisqually basin, of which Tahoma Creek represents 10% as measured by area. Surveys of the delta completed in 1945, 1956, 1985, and 2011 provide estimates of total deposition over multiple time periods (Czuba et al., 2012b). We convert these total loads into bed load estimates with the assumption that bedload represents 20% of the total volume, a reasonable value for steep environments (Pratt-Sitaula et al., 2007; Wallick et al., 2010). We compare these estimates of bed load to the volumes of bed load predicted over the same time periods within Tahoma Creek using our thresholded rating curve and synthetic hydrology. As sedimentation rates from 1945 to 1956 are dominated by input from the 1947 Kautz Creek debris flows, it is excluded from this comparison. The results, shown in table 1, predict that Tahoma Creek provides between 30% and 50% of the total bed load deposition within Alder Lake.

#### 4.5.3. Sediment Budget for the Upper Nisqually Basin

Using the predicted coarse sediment transport from Tahoma Creek, and a rough sense of the relative contributions from the Tahoma, Kautz and upper Nisqually basins derived from repeat LiDAR between 2008 and 2012, it is possible to construct a sediment budget for the major pro-glacial basins within the contributing area of Alder Lake. Assuming that Tahoma Creek and the Nisqually River export similar volumes of coarse sediment, while Kautz Creek exports very little, the three basins combined are estimated to account for 85% to 95% of the coarse material deposited within Alder Lake, with the remaining 5% to 15% coming from non-glacial areas. Note that, with the exception of the non-glacial contribution, these values were derived entirely from repeat LiDAR, independent of the total accumulation rate measured downstream, yet provide a near exact match. Our estimate of the non-glacial sediment contribution translates to a production rate of  $100\pm50 \text{ kg/km}^2/\text{yr}$ . A lower delta within Alder Lake, with an entirely nonglacial source area, has also be re-surveyed, and accumulated sediment at a rate of  $74\pm24 \text{ kg/km}^2/\text{yr}$ , in close agreement with our estimates ((Czuba et al., 2012b)).

# 4.5.4. Mt. Rainier Bedload Measurements

A limited number of sediment transport measurements taken on rivers draining Mt. Rainier provide some means of directly assessing our sediment rating curve. Compiled measurements from multiple streams, seen in Czuba et al. (2012a), are fit with a power law, giving an exponent of 2.36 and a coefficient of 5.7. Extracting just the White River data, which is the most abundant from a single source, gives an exponent of 3.16 and a coefficient of 0.2. Comparing our results from the un-thresholded rating curve, our coefficient is significantly higher, but our exponent seems to be roughly similar to these measurement. However, these measurements were taken at positions significantly lower in their basins that Tahoma Creek, and occurred over flows ranging from 4% to 12% of the mean annual flood for the sampling sites, and so represent only the lowest end of the rating curve.

# 4.6. Summary

Our estimates of total loads observed over the 2002-2012 time period, and the sediment rating curve derived from these total loads, appear to be reasonable. The close agreement between bed load deposition in Alder Lake and the predicted bed load transport from Tahoma Creek is particularly heartening, though the lack of a measured grain-size distribution for the Alder Lake sediments means this target is subject to its own uncertainty. The percentage of coarse sediments contained within these delta deposits was assumed to be 20%, but could easily vary by a factor of two. However, even given this range, the estimated mean annual bed load from Tahoma Creek would represent between 20% and 80% of the coarse deposition within Alder Lake. That our values fall entirely within this physically possible (and plausible) range represents a distinct success when considering the orderof-magnitude uncertainties that are common in bed load estimates. Using information about Tahoma Creek transport, and the relative contributions of the Tahoma, Kautz and upper Nisqually basins, results in sediment budget for the larger Nisqually basin that nearly exactly matches what is measured in Alder Lake. Estimates of non-glacial sediment production rates based on these values are very close to those independently estimated through repeat

surveys of a second, non-glacially sourced delta within Alder Lake. Taken together, the coherency of the sediment budget for the Nisqually basin, developed through morphologic budgeting methods and independently verified through Alder Lake sedimentation rates, provide a strong assurance that the methods presented above accurately reflect coarse sediment transport within Tahoma Creek.

#### 5. Evaluating Bedload Transport Equations

The analysis above provides a means to assess the performance of bedload transport equations in these settings, both in terms of the total loads they predict and the form of the sediment rating curves they produce. Sediment transport equations are subject to significant uncertainty, but remain a common element in stream analyses, owing to the difficulty and expense of a full bed load measurement campaign. Many equations have been shown to be particularly inaccurate in steep settings, where they often over-predict transport by several orders of magnitude (Bathurst, 2007; Yager et al., 2007; Chiari et al., 2010; Nitsche et al., 2011). Here, we choose to test three equations - Parker, surface-based (1990), Wilcock and Crowe (2003), and Recking (2010). While a more recent formulation of Recking's equation is presented in Recking (2013), in which the original piece-wise formula is simplified using a logarithmic-matching technique, we elect to use his original formulation because it allows for the direct manipulation of the critical shear stress. In addition to the original forms of the transport equations, we also run them using one of two modifications that aim to reduce the tendency to over-predict transport. In one case, we run all equations with a reduced slope, accounting for energy losses due to large roughness elements. We operationalize this by combining equations (31) and (32) in Rickenmann and Recking (2011) to get

$$S_0 = p^e S^{1-ez}$$

where S is the original slope,  $S_0$  is the reduced slope, and p, z and e are empirical constants. We set p = 0.07, z = 0.47 and e = 1.2, values suggested in their paper that provide a reasonable fit to their data as well as the compiled data of Palt (2001) and Chiari et al. (2010).

In the second case, we replace each the originally specified critical shear stress of each equation with the slope-dependent formula of Mueller et al. (2005)

$$\tau_{*c} = 0.21S + 2.18$$

derived from field observations in Idaho streams. Recking's formula was also based on this data, and so his formula for critical shear is similar to that of Mueller (figure 14).

# 5.1. Inputs and Methods

Inputs to these formulas include topography, sediment size distributions, and estimates of hydraulic parameters. Cross sections and slopes were derived from 2008 LIDAR. To estimate grain-size distributions in Tahoma Creek, 25 surface and 6 subsurface samples were taken along the lower 5 RKs. Surface samples consisted of 200-clast Wolman pebble counts, recorded at half-phi intervals down to 2mm. For subsurface samples, the surface layer was first removed, and 300-500 kg's of gravel was collected. Particles coarser than 32mm were sorted and weighed in the field, while a fine split of  $\sim 5 \text{ kg}$ of the <32mm material was retained and later measured in the lab down to 0.35mm. Equations were run using a combined surface/subsurface distribution. For the Parker equation, this distribution was truncated at 2mm and the relevant sediment parameters taken from this new distribution, as per the original paper. Flow depths and velocities at a cross section were estimated using Ferguson (2007) flow-resistance equation, which provided good agreement with stage-discharge relations observed at the bridge; it also received high marks from Rickenmann and Recking (2011) when tested against a large field dataset. Each cross section was divided into subsections defined by sequential station-elevation points, and transport calculated for each subsection at a given flow. This process accounts for the significant lateral variability in shear stress, providing a more accurate measure of transport (Ferguson, 2003). The subsections were then summed to give total transport across the section. We used this process to calculate sediment rating curves between 0.3 cms and 80 cms at 50 locations, spaced 100m apart down the lower five kilometers of the channel. These rating curves, in combination with the synthetic flow records described in section 4.1, were then used to predict total loads over the time steps for which LIDAR-derived estimates



Figure 11: sediment transport predictions from transport equations. A) Downstream trends for unmodified equations of Parker (1990), Wilcock and Crowe (2003) and Recking (2010). B) Box plots of transport predictions over downstream extent. LiDAR-derived values are shown as black lines

exist. This process was repeated for all three equations, using three forms (unmodified, reduced slope, and Mueller's  $\tau_{*c}$ ), for a total of nine results.

#### 5.2. Results

All three equations and their modified forms did a reasonable job of reproducing the observed consistency in downstream transport rates, albeit with significant noise (figure 11). The unmodified Parker and Wilcock-Crowe equations show a minor decrease in transport rates moving downstream, suggesting an aggradational state, while the unmodified Recking equation shows a slight increasing trend, suggesting an incisional state. Both modifications applied to these equations reduce transport as a function of slope, such that the modified Parker and Wilcock-Crowe equations show downstream trends similar to that of the Recking equation. However, these trends are minor compared to the point-to-point variability. As such, total loads for each equation were averaged in the downstream direction and compared to similarly averaged transport as estimated from LiDAR. The predicted transport volumes vary over five orders of magnitude for a given time period, highlighting the immense uncertainty in predicting bed load transport. For 2002-2008, both the unmodified Parker equation and the Wilock-Crowe equation modified using Mueller's  $\tau_{*c}$  produce reasonably accurate results. However, these apparent agreements vastly overstate the success of these equations. As seen in figure 12, an unrealistic percentage of the predicted transport for these equations occur during very low flows. In the case of the Wilcock-Crowe equation, over 80% of the predicted 2002-2012 load was moved during flows below 5.5 cms, our field-estimated critical discharge. For this equation, as well as the Parker equation, the 2006 flood barely registers. Similar results can be seen by comparing the rating curves used to generate the total volumes (figure 13); both equation over-predict of transport at lower flows, but come close to the correct total loads as a function of the under-prediction of transport during high flows, and particularly during the 2006 flood.

For the 2008-2012 period, the unmodified Recking equation and the Muellermodified Parker equation predict total volumes is rough agreement with those



Figure 12: Cumulative transport for the 2002-2012 period, expressed as a percentage of the total transport predicted.

observed. The rating curves for these equations sit reasonably close to the un-thresholded LiDAR rating curve, particularly over the range of lower flows that produced most of the 2008-2012 transport, providing some confidence that their success is physically meaningful.

Taken as a whole, it is clear that none of the equations, or their modified forms, provides a particularly accurate estimate of sediment transport processes in Tahoma Creek. Of the two modifications applied, the use of a modified critical shear stress did a reasonable job of improving transport predictions over low to moderate flows, while the use of a reduced slope caused significant under-prediction of transport rates at all flows. The good agree-



Figure 13: Bedload transport rating curves at RK 1.3, as compared to LiDAR derived curves. Solid lines represent unmodified equations, while dashed lines represent the use of Mueller's  $\tau_{*c}$ . Grey shading indicates  $2\sigma$  uncertainty range for our LiDAR-derived results, as quantified through a Monte Carlo analysis. Rating curves using the shear-partitioning method plotted very low, and so were omitted for clarity.

ment between our field-estimated critical shear stress and those predicted by Mueller's equation gives some confidence that the improvements seen using this modification are the result of an improved characterization of processes within Tahoma Creek, and not simply a numerical fluke.

# 6. Discussion and Conclusions

The methodology we present here uses repeat aerial LiDAR to measure bedload transport and interpolate a rating curve that describes daily trans-



Figure 14: Comparison of critical shear stress values from a variety of equations to field estimated values for Tahoma Creek. Grey bars indicate the range of slopes and critical shear values seen near the Tahoma Creek Bridge. Field values for critical shear were estimated by observing what level of flows generally produced shifts in the rating curve.

port rates within Tahoma Creek. Uncertainties, both quantifiable and unquantifiable, accumulate as we move from first-order estimates of volumetric change into second-order estimates of transport, and again when we interpolate a rating curve based on these transport estimates. While the total uncertainty may be sizable, several lines of reasoning show that our results are plausible and provide reasonable estimates of bed load transport in Tahoma Creek. Most notably, direct measurements of the total loads transported by the Nisqually River, derived from repeat surveys within Alder Lake, closely match our estimates of transport in Tahoma Creek, even when accounting for the uncertainty introduced when converting the measured total loads into bed load.

Several aspects of Tahoma Creek make it particularly well-suited to the morphologic budgeting techniques we use here. The basin is very active, with clearly defined and largely vegetation-free zones of transport. Only a small percentage of the active channel is submerged during the later summer months, leaving it readily surveyed using conventional airborne LiDAR. Our ability to translate volumetric change into coarse sediment loads was dependent on surveys that encompassed the entire basin, and was made significantly more accurate by the localized nature of the source areas. Our ability to back out a sediment rating curve from our observations of transport was aided by the immense disparity in the hydrology and accompanying sediment transport during the two time periods. Given that the rating curve analysis boils down to a best-fit procedure in which, in essence, a line is fit to two fuzzy points, the greater the separation between these points, the less uncertainty there will be in the results.

Even with these benefits, our estimates of the 2008-2012 transport, shown in figure 7, involve a 1- $\sigma$  uncertainty of nearly  $\pm 50\%$ . Such uncertainties suggest that, at present, applying similar techniques in vegetated or relatively inactive areas may be difficult. Beyond just the physical characteristics needed, there are also significant logistical costs involved in these methods. Aside from the (sizable) costs of acquiring, processing and analyzing these data, enough time must elapse between surveys such that the accumulated change can be meaningfully detected by the chosen survey methods. The 10 years of change documented here illustrate that, even in active settings, this may be a significant period of time. Fortunately, given the rapid pace at which LiDAR is becoming a mainstay in a variety of fields, the technology will continue to improve while costs continue to fall. This will decrease the investment involved with obtaining repeat surveys, as well as increasing the probability that historical data may cover a given site of interest. New technologies that improve our ability to resolve vegetated or submerged topography would increase the applicability of these methods.

For the reasons listed above, the full methodology we have presented here seems unlikely to become the norm for quantifying sediment transport. However, these methods provide estimates of transport during high flows that are virtually unmeasurable otherwise. Importantly, the uncertainty of our estimated transport rates is independent of that rate. As a result, we have a significantly higher relative confidence in our estimates of transport during the 2006 flood than for any other flow. This is apparent in figure 13, where both LiDAR-derived curves converge at high flows, situated at the same location that uncertainty within these curves is a minimum.That this method excels exactly where most struggle should make it a valuable tool when used in combination with other methods.

We also note that the methodology we use here to derive sediment rating curves for Tahoma Creek is in no way tied to repeat LiDAR surveys or bed load. Similar methods could be applied equally well in any situation where sediment loads, including suspended or total loads, were known for at least two time periods, along with records of discharge for the intervening period. Repeat reservoir surveys are the most obvious target, especially given the increasing ease and accuracy of such surveys.

# 6.1. Statistical Analysis of Topographic Data

The defining trend of modern survey techniques has been the increasing density of information. Laser-surveying techniques have the ability record millions of xyz points in a relatively short period of time, defining features over scales ranging from millimeters to kilometers. While the products of these technologies are qualitatively similar to previous topographic representations, the shear quantity of information now available requires new tools and methods to process, analyze and interpret this data. Crucially, we need to develop measures of uncertainty and identify sources of errors inherent in these new products before we can fully exploit them. In this regard, while the number of observations may be daunting, they also open up new opportunities for statistical methods to shine. Much as statistical mechanics can describe the aggregate properties of countless particles with arbitrary accuracy, applying these techniques to high-density topographic datasets may allow for precision at levels well below the accuracy of individual measurements. Such methods underlie the terrain-matching process we use here to produce accurately aligned datasets, and are at the heart of our analysis of cell-level and alignment uncertainty.

### 6.2. Sediment Transport Prediction

Our LiDAR-derived rating curve was used to assess the performance of several modern sediment transport equations. Both the Parker and Wilock-Crowe equations over-predict transport rates for low flows, often by several orders of magnitude. We find that this tendency to over-predict transport is best explained by the overly low critical shear stress values used in both equations. Replacing the originally specified values with Mueller's slopedependent formula for critical shear stress improves the agreement for both equations, though it is still not strong. The Recking equation, which uses a formula for critical shear stress similar to that of Mueller, shows reasonable agreement with the LiDAR rating curve in these lower flows using either formulation. Field estimates of the critical shear stress for Tahoma Creek closely match values predicted from both Mueller's and Recking's equations, providing some confidence that this improvement gets it right for the right reasons (figure 14). Accounting for form-drag through the use of a reduced slope, a common strategy when predicting transport in step-pool systems, leads to significant under-predictions of transport across the full range of flows.

In contrast to the low flow regime, all transport equations significantly under-predict transport rates during large events such as the 2006 flood. As noted above, LiDAR-derived transport rates at these flows are likely to be close to correct even if the broader rating curve is not. Instead, we suggest that this under-prediction is due to the lack of available transport observations during high flows on which to base models. To our knowledge, the highest bed load transport rate ever measured was 3.9 kg/m/s, taken on the Toutle River in aftermath of the the 1980 Mt. St. Helens eruption. We estimate here that, at the peak of the 2006 flood, transport rates were close to 400 kg/m/s. The volume of sediment transported during the three days encompassing this flood constitute a significant percentage of the total coarse load exported by Tahoma Creek over the past 60 years. Failing to properly account for these large events may lead to a significant underestimation of sediment loads when considered over decadal time-scales.

# The geomorphic impacts and historical precedence of debris flows within Tahoma Creek, Mount Rainier, WA

Scott Anderson, John Pitlick and Paul Kennard

### Abstract

Paraglacial sedimentation processes acting within zones of recent glacial retreat have the potential to drive aggradation in rivers throughout the Pacific Northwest, further exacerbating flood hazards that may be independently intensifying as a function of warming. Mt. Rainier has become the focal point of research into these processes, as both Mt. Rainier National Park and large population centers downstream sit directly adjacent to active fluvial systems. In this work, we focus on recent geomorphic change within Tahoma Creek, a pro-glacial basin on the southwest flank of Mt. Rainier. Over 27 debris flows have been recorded within the basin since the late 1960s, significantly more than any other basin around Mt. Rainier. Repeat aerial LiDAR, flown in 2002, 2008, and 2012, indicate that debris flows have mobilized over  $2.6 \times 10^6$  m<sup>3</sup> of recently exposed pro-glacial sediments within the South Tahoma Glacier fore-field. Of this, a little over half has been deposited within the upper basin, while the remaining half transited the lower fluvial reaches to exit the system. These lower reaches have remained vertical stable over this period. The age and location of young alders growing within

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the active channel indicate that debris flows translate into increased channel mobility within the lower basin, but manifests primarily as lateral motion, with only minor associated aggradation. This state persists for several years after debris flow activity ceases. An analysis of historical records back to 1915 suggest that the channel is functioning much the same now as it has over the past century. Dendrochronologic methods were used to construct a record of debris flows within the basin extending back to 1500. This record shows that a suite of debris flows, similar in extent to those of the recent decades, occurred between c.1840 and c.1870, following the onset of retreat out of the Little Ice Age. While records are sparse, all debris flows recorded before 1840 also coincide with the onset of glacial retreat, as evidenced by moraine stabilization dates. This correlation between the onset of retreat and the occurrence of debris flows could plausibly be related to increased sediment availability or higher frequencies of outburst floods, which have been common triggers of the modern debris flows. Taken together, there is no clear evidence that recent warming or the frequency debris flows have pushed Tahoma Creek outside of its historical range of variability. However, Tahoma Creek may be sufficiently steep to act as a conduit for increasing sediment loads, leaving downstream communities at risk.

# 7. Introduction

Over the past 40 years, dozens of debris flows have swept down the flanks of Mt. Rainier, WA, transporting millions of cubic meters of unconsolidated sediments from proglacial settings into downstream valleys (Driedger and Fountain, 1989; Walder and Driedger, 1994a; Copeland, 2009). These debris flows constitute one part within a suite of geomorphic processes that act to mobilize sediments left in the wake of glacial retreat, collectively referred to as paraglacial processes. This term was originally coined in the context of elevated sediment loading that occurred in the millennia following the last glacial maxima (LGM), which has sometimes persisted into the modern era (Church and Ryder, 1972; Hallet et al., 1996). However, applying this concept to the complex sequence of advances and retreats of the past centuries is not straight forward (Orwin and Smart, 2004). The relative importance of stream flow and sediment availability in driving channel morphology over sub-millennial time scales may vary significantly, both regionally and within a single watershed, such that both periods of glacial advance (Hart et al., 2010; O'Connor et al., 2001) or retreat (Church and Ryder, 1972) may drive downstream aggradation. Understanding the relative importance of these factors is then crucial when considering the possible geomorphic responses to recent climate change.

In Washington State, glacial retreat and a changing hydroclimatology may be working in concert to elevate sediment loads above the already high rates conditioned by the wet climate and volcanically active terrain. Glaciers on Mt. Rainier have retreated significantly over the past century (Nylen, 2004), and examinations of recent debris flows have shown that nearly all have initiated within recently exposed glacial sediments (Walder and Driedger, 1994a; Copeland, 2009). While less well-documented, there are indications that precipitation and flooding have become more intense in recent decades, potentially accelerating the rate at which sediment is being mobilized and transported downstream (Mass et al., 2010). When aggradation occurs within a stream reach, the stage for a given flow increases, elevating the likelihood of damage associated with flooding. Aggradation also predisposes rivers to lateral motion, further compounding difficulties for flood hazard management. A thorough discussion of these processes and their potential impacts for the lowland regions downstream of Mt. Rainier is presented in Czuba et al. (2012a).

This broad concern was brought to the forefront in the wake of a massive 2006 flood, with regional impacts. Dozens of debris flows were recorded through out the Cascade Range, including multiple such events on Mt. Rainier (Copeland, 2009), Mt. Adams (Williams, 2011), Mt. Hood (Ellinger, 2010), and Mt. Jefferson (Sobieszczyk et al., 2008). Damages to Mt. Rainier National Park infrastructure totaled over 36 million dollars, and forced the park to close for six months. To date, however, geomorphic and hydrologic evidence for persistent trends toward more energetic events remain ambiguous. Crucially, there is rarely a strong handle on the frequency or intensity of flooding and debris flows before the early years of the 20th century, and, particularly in the case of debris flows, such records are often incomplete. Lacking this baseline data, it is difficult to discern whether or not these recent events sit outside the historical range of variability for these systems.

### 8. Overview

This research takes a multi-faceted approach in trying to understanding the process linkages between glacial dynamics, debris flows and the downstream fluvial response on the flanks of Mt. Rainier, and attempts to provide some historical context for recent events. We focus our attention on Tahoma Creek, a single watershed on the southwest flank of Mt. Rainier. Tahoma Creek has experienced at least 27 debris flows since 1967, significantly more than any other basin around the mountain. These debris flows have left a very visible impact on the upper reaches of the basin, and have forced Mount Rainier National Park to limit access to much of the western margin of the Park. Downstream, the channel has required repeated dredging to ensure adequate freeboard below the Tahoma Creek Bridge, a span along the Park's main access highway. There remains concern that aggradation along this reach has the potential to destroy this structure.

Analyzes are broken into three distinct sections, encompassing increasingly longer time-scales. Section one covers an analysis of the modern channel change following recent debris flows, using both LiDAR datasets flown in 2002, 2008 and 2012, and dendrogeomorphic methods applied to alder stands growing within the active channel. Section two entails a coarse analysis of geomorphic change over the past century based on the analysis of historical records. These records include aerial and oblique photos, bolstered by dendrochronologic estimates of surface ages within the active channel. Section three attempts to place the modern debris flows within a broader context through a c. 500-year dendrochronologic reconstruction of debris flows and flooding. The resulting chronology is then compared to reconstructions of climate and glacial mass balance to examine which, if any, of these factors may control debris flow frequency. Finally, these analyses are considered together to provide a more complete picture of sediment fluxes, both specifically within Tahoma Creek and in similar drainages around Mt. Rainier.

#### 8.1. Regional Setting

Mt. Rainier is the largest member of the Cascade Range, a volcanic arc running north-south within the Pacific Northwest. The mountain is home to 26 glaciers, comprising over a cubic mile of ice (Driedger and Kennard, 1986). It is geomorphically active, a result of the steep terrain, relatively weak rock, and the combined impacts of glacial and volcanic processes. Eruptive events over the past millennia have produced massive lahars, the deposits of which form valley floors in many drainages (Scott et al., 1995). Between these events, a more steady stream of rockfall, debris flows and floods continually move mass down into lower valleys, ultimately making its way into the Puget Sound (figure 15).

Glaciers throughout the Cascade Range, like glaciers throughout the world, have been retreating rapidly since the mid-19th century (Nylen, 2004; Lillquist and Walker, 2006; Kovanen, 2003), though this retreat has been punctuated by stand-stills or slight re-advancements in the early 1970s and 1990s. Glaciers throughout the range advance and retreat in response to variability in both summer ablation rates, a function of temperature, and winter accumulation rates, primarily a function of precipitation (Burbank, 1982).



Figure 15: Regional map of study site with blowup of the southern flank of Mt. Rainier. A and B text denote locations of photos seen in figure 16

# 8.2. Tahoma Creek

Tahoma Creek drains  $38 \text{ km}^2$  of the southwest flank of Mt. Rainier, with its headwaters in the South Tahoma Glacier (figures 15, 16 and 17). Since its Little Ice Age (LIA) maximum in 1840, this glacier has retreated over three kilometers (Sigafoos and Hendricks, 1972), leaving behind extensive sediments and stagnant ice within the upper basin. Hillslope failures along



Figure 16: Photographs of Tahoma Creek. Locations of photos are indicated by text in figure 15

these moraines provide the majority of sediment contained within frequent debris flows. Once initiated, these debris flows generally transit through a bedrock-confined reach between river kilometers  $(RK)^4$  8 and 10, depositing much of the entrained material within the wider valley between RKs 5 and 8. Below this reach, fluvial processes dominate. The lower channel is primarily single-threaded, though mild braiding is not uncommon. The wetted channel is usually less than 10 meters wide during normal flows, and sits within a larger active channel 30-70 meters wide. The active channel is composed of bare gravel that supports stands of red alder, and is bordered by a 2-3m high floodplain/terrace surface, consisting of both fluvial overbank

 $<sup>^4 \</sup>rm Valley$  locations are noted as kilometers upstream from the Tahoma Creek Bridge. The bridge then sits at RK 0, with the 2012 terminus located near RK 13 (figure 17)
sediments and lahar deposits. These elevated surfaces support extensive conifer forests, many of which are in a late-successional stage and include individuals over 1,000 years of age. Since 1967, over 27 debris flows have been documented within the Tahoma Creek basin, concentrated within two periods in 1967-1972 and 1986-1992 (Crandell, 1971; Walder and Driedger, 1994a). These events have occurred exclusively in the late summer and early fall. The later of these high-frequency periods was well-documented by Joe Walder and Carolyn Driedger, research scientists with the USGS Cascade Volcanoes Observatory. In their 1995 report, they propose that outburst floods from South Tahoma Glacier initiated the majority of the documented debris flows. This claim was based on the observation of large volumes of water issuing from the upper basin within the debris flows, indicating storage somewhere within this reach. Statistical analysis of temperature and precipitation records immediately preceding debris flows showed that these events occurred preferentially during either particularly warm, dry periods or large rainfall events. This result countered previous reports suggesting that geothermal melting occurring at the base of South Tahoma Glacier was responsible for the large volumes of water observed. However, there remains scant geomorphic evidence for these outburst floods, leaving the door open for other explanations. The most plausible alternate location for water storage in the upper basin is within saturated sediments, intermixed with remnant ice, sitting in lateral moraines.



Figure 17: Map of Tahoma Creek and surrounding area. Valley distances in kilometers above the Tahoma Creek bridge are marked. Contour interval is 100 meters. TG is the Tahoma Glacier, STG is the South Tahoma Glacier. Major roadways are indicated by solid black lines - the Westside road closure is indicated by the transition to a dashed line.

### 9. Channel Response to Modern Debris Flows

Understanding the downstream response of Tahoma Creek to recent debris flows is a first-order priority for the Park Service, given concerns about the fate of the Tahoma Creek Bridge. More broadly, the timing and extent of this downstream response provides some indication of how large a disturbance the upstream sediment loading represents for the system as a whole. Changes since 2002 are well documented in several aerial LiDAR datasets, which provide a detailed look at geomorphic activity over the entire basin at the meter-scale. For activity prior to 2002, we use the age and locations of alder stands growing within the active channel to infer periods of increased channel activity and aggradation.

## 9.1. Repeat LiDAR Analysis - Methods

Tahoma Creek is covered in three aerial LiDAR datasets, flown in December, 2002, September 2008 and August, 2012, targeting leaf-off, snow-free conditions. The later two datasets cover the entire basin from glacial source to the confluence with the Nisqually River, while the 2002 dataset extends from the source down to RK 1.3, missing the lower two kilometers of the valley. The first step in using these datasets to measure change within Tahoma Creek is ensuring that they are accurately co-referenced. A detailed discussion of these datasets and the steps taken to co-reference them is presented in the previous chapter, and so we present only a summary here.

# 9.1.1. Co-referencing

Change analysis only depends on the relative geo-referencing of the datasets, and so we arbitrarily chose the 2008 survey as our ground truth, and aligned the remaining datasets to it. All datasets were projected into NAD83 UTM coordinates, using the NAVD88 vertical datum. The 2002 and 2012 rasters were snapped to the 2008 grid, providing exact cell overlap. Horizontal offsets between the surveys were detected and removed using a terrain-matching method similar to that presented in Delong (2012), while the vertical alignment was assessed by comparing the compact, level surfaces of several roadways captured in the surveys. Once properly aligned, sequential DEM's were differenced to create two DEM's of difference (DoDs), one encompassing change from December 2002 to September 2008, and one encompassing change from September 2008 to August 2012.

### 9.1.2. Analysis Methods

Both LiDAR analysis and field inspections indicate that the forested valley sidewalls did not contribute any significant volume of sediment to the basin between 2002 and 2012, and so we restricted our analysis of change to the active channel and adjacent floodplain/terrace surfaces. These areas were subdivided into 100m sections, as measured along the valley centerline, and the net volumetric and mean vertical change was calculated for each section. Since our surveys encompass the entire basin, and presumably sediment only leaves the basin through down-valley transport, we translate the net volumetric change upstream of a point into the volumetric flux of sediment past that point. Suspended sediment is predominately sub-glacially sourced (Fahnestock, 1963; Mills, 1979), and quickly exits the basin with minimal geomorphic impact, and so we take this volumetric flux to represent bed load, as opposed to total load. A comparison of the grain size distribution within moraines around Mt. Rainier to the bed material of Tahoma Creek supports this assumption (Mills, 1978). A more complete discussion of the assumptions implicit in this method of measuring bed load are presented in Anderson and Pitlick (2013). Figure 18 shows the results of this analysis - blue lines show mean vertical change every 100m downstream, while black lines show our estimation of transport through the basin, encompassing both debris flow and bed load transport. The uncertainty within our estimates of sediment transport are quantified by the addition or subtraction of a vertical offset across the entirety of the derived DoDs, representing uncertainty in the vertical alignment of the parent surveys.

#### 9.1.3. Observed Change, 2002 to 2008

The observed change from 2002 to 2008 is dominated by the impact of the November 2006 flood, with an estimated peak discharge of 68 cms in the lower reaches of Tahoma Creek. 2.3 million  $m^3$  of material was exported from the upper basin, predominately sourced from broad hillslope failures along lateral moraines. Approximately one million  $m^3$  of this material was deposited along the valley floor between RKs 5.5 and 9. The remaining 1.3



Figure 18: Vertical channel change and volumetric sediment transport, as measured from LiDAR. The 2008-2012 transport rates are also plotted on the 2002-2008 scale to better illustrate the disparity between the two periods. Uncertainty in the transport estimates are quantified by assuming a 1 uncertainty of 2.5 cm in regards to the vertical alignment of the parent surveys.

million m<sup>3</sup>, along with 50,000 m<sup>3</sup> entrained from the lower valley, transited the fluvial reach to exit the lower extent of the 2002 survey. The majority of this sediment likely continued downstream to enter the Nisqually River. The fluvially-dominated reach below valley km 5.5 was, on average, weakly incising over this period, though deviations exist. A spike of aggradation occurred around RK 4, likely the result of the river flowing through a recently cleared forest, with numerous snags acting as large roughness elements. The aggradation around valley km 3 is the result of a distinct break in the valley slope, dropping from 0.06-0.07 to 0.03-0.045. Field inspection shows that deposition in this area is common, and likely represents a zone of temporary storage for sediment transported during high-flow events.

## 9.1.4. Observed Change, 2008 to 2012

The period from 2008 to 2012 was notably less active than the prior one - only  $250,000 \text{ m}^3$  of material was removed from the upper basin, largely sourced from sediments sitting on a steep bedrock step just below the 2012 terminus. Along much of the length of the lateral moraines, sediment moved downslope, recharging sediment stocks sitting on the channel floor. However, very little of this material was further mobilized downstream. Deposition of material began near RK 8.5 and extended down to RK 7. Below this, transport volumes were essentially constant at 100,000 m<sup>3</sup>, accomplished through fluvial processes. While channel restructuring occurred in the lower reaches, there was essentially no net vertical change. The reach at and below the bridge provide the one exception, aggrading roughly half a meter. This localized aggradation may be a response to vertical changes along the Nisqually River, which in turn may be backing up on the large sediment loads from Tahoma Creek. While the discharge along the Nisqually is significantly higher than that of Tahoma Creek alone, Tahoma Creek is nearly twice as steep as the Nisqually at this confluence. This may partially explain why aggradation under the bridge has been so persistent while the majority of the upstream channel has remained vertically stable.

Relative transport over the two time periods was similar. In both cases, roughly half of the total load mobilized from the upper basin was deposited within the valley above RK 5, while the remaining half transited the lower fluvial reaches to exit the system. Over the entire period of record from 2002 to 2012, the fluvially-dominated zone below RK 5 remained vertically stable, despite the large volumes of sediment that transited through these reaches. Localized instances of aggradation appear to be related to specific valley controls, and do not suggest a broad aggradational state. Sediment waves, which have been an item of interest within the Park for their potential to exacerbate local flood hazards and channel mobility, are not evident, though such features may not be readily apparent when examining change integrated over several years.

### 9.2. Alder Establishment - Methods

The active channel of Tahoma Creek is dotted with stands of red alder (*Alnus rubra*) growing on sandy or bare-gravel surfaces. These stands establish when a surface remains stable long enough for seeds to germinate and saplings grow such that they can withstand subsequent flooding. The age structure and location of these trees should thus contain information about the historical activity of the channel. To this effect, cores were taken from 63 alders within 21 distinct stands in the lower 5 RKs to determine establishment dates. Two cores were taken from each tree 30 to 50 cm above

the ground surface. These cores were mounted and sanded, and the rings counted. If the pith was not cored, the number of missing rings was estimated. Given the small diameter of these alder, most trees included at least one core with pith.

All sampled trees reached coring height between 1961 to 2003, with the majority after 1980. The ecesis interval for these trees, representing the time lag between surface stabilization and the tree achieving coring height, is estimated to be a single year. This is based on the rapid colonization and growth of alder saplings seen on 2006 debris flow and flood surfaces. The date of stabilization for a surface was taken as the oldest establishment date in a stand (pith age plus one year ecesis). To explore the vertical position of these stands, they were first digitized as polygons using 2008 LiDAR and 2009 aerial photography, and the mean elevation of the stand measured from 2008 LiDAR. This elevation was then expressed as a height above the adjacent 2008 low-flow water level.

#### 9.3. Alder Analysis - Results

Alders growing within the active channel show a cluster of establishment between 1989 and 1995, with a distinct peak in '93-'94 (figure 19). This time period matches that of debris flows recorded from 1986 to 1992 with a threeyear lag. The peak in '93-'94 coincides with both the end of recorded debris flows, as well as three consecutive years of low peak flows as recorded at the National Gage on the Nisqually River. As such, this peak could plausibly



Figure 19: Results of alder core analysis. Establishment dates, taken as pith age plus one year ecesis interval, of all cored individuals are shown in panel B. overlain are records of modern debris flows. Annual peak flows as recorded at a USGS gage on the Nisqually River, 27 kilometers downstream of the Tahoma-Nisqually confluence, are shown to illustrate the connection between flood hydrology and the peak in alder establishment in '93-'94. Stand elevations are shown in panel C, plotted in relation to the 2008 active channel.

be a function of reduced sediment loading, the absence of channel-altering flows, or both. Given the relatively high number of establishments occurring during the preceding years, and the lack of similar spikes during other low flow years, this last explanation seems most likely; increased sediment loading and channel activity following debris flows created elevated or distal surfaces wellsuited to colonization, while a series low peak-flow years allowed seedlings on these surfaces time to fully establish. Whatever the exact explanation for the '93-'94 spike, debris flows and alder establishments are well correlated, indicating that these events did have an appreciable downstream impact. A similar, though muted, cluster of establishments occurring between 1976 and 1978 could be linked to the pulse of debris flows occurring between 1967 and 1971, with a lag of 5 to 8 years.

Alder stands were even-aged, though in several instances, multiple age classes co-existed at a single location. These stands were generally situated on slightly perched or protected surfaces along the active channel. However, partially buried trees and those with adventitious roots attested to the fact that many stands remain within the vertical range of recent channel activity. Using the 2008 LiDAR to quantify the vertical position of these stands relative to the low-flow water level, most are seen to sit near the 95th percentile in the local distribution of elevations. This holds for stands established both in the years following the most recent debris flows and for those established prior. Only near RK 2 do stands sit notably above the modern active channel, supplementing evidence below that this reach is particularly prone to aggradation. While these measurements are presented simply as comparisons to the 2008 channel, the LiDAR analysis above indicates that there has not been significant vertical channel change over the 2002-2012 period, providing some confidence that the 2008 channel is a reasonable baseline for comparing the recent activity with disturbances several decades ago.

Taken as a whole, the age structure and channel locations of these alders suggest that recent debris flows in the upper basin have translated into increased channel activity in the lower basin, represented by the marked increase in alder establishments during the years following these events. This increase is inferred to have manifested predominately as increased lateral mobility, with only moderate associated aggradation. This period of increased activity appears to have ceased several years after the debris flows themselves, providing a rough estimate of the time needed for Tahoma Creek to process the newly deposited sediment in the upper basin.

## 9.4. Summary of Modern Change

Taken together, the analysis of both aerial LiDAR surveys and alder stands present a relatively coherent picture of the downstream fluvial response to debris flows in the upper basin. These events do increase channel mobility, and so presumably total sediment fluxes, but do not produce significant aggradation. This suggests that the channel is sufficiently steep and energetic such that increased sediment fluxes can be accommodated through changes in local bed texture or structure, while remaining vertically stable. Increased mobility may persist for several years following debris flows, as the stream reworks the newly deposited sediment. However, these periods do not appear to persist longer than a decade, and may be significantly shortened by large floods. Only a small percentage of the total depositional volume is likely to be remobilized, as the channel becomes inset within the deposits and armored by the coarser fraction of the debris flow sediments. Much of this sediment will likely remain in long-term storage within this wide valley-floor between RKs 5 and 8.

Aggradation under the bridge may be more a function of changes along the Nisqually River, though this in turn may be driven by sediment loads from Tahoma Creek. Regardless, the persistent need to dredge sediment from beneath the bridge does not appear to reflect a basin-wide aggradational trend.

### 10. Analysis of Historical Records

Historical records of Tahoma Creek extend back nearly to the Park's establishment in 1899. There is a rich aerial photo record beginning in 1951, providing a measure of channel activity over the past decades. Various surveys and oblique photos exist back to 1904, though these early records are almost exclusively focused on the bridge and immediately adjacent channel, and so provide a relatively limited view of these early years. During the course of the dendrochronologic analysis presented below, several relatively young surfaces were identified within the lower active channel; the age of trees growing of these surfaces were used to place an upper limit on when they were last active. Taken together, these records provide some context for the recent geomorphic activity and a better understanding of what factors control downstream activity.

## 10.1. Aerial Photo Analysis

Aerial photographs of Tahoma Creek were obtained from Park records for the years of 1951, 1960, 1969, 1979, 1984, 1989, 1996, and 2009. The 2009 imagery was provided as a georeferenced digital image, while all earlier records were stored as physical prints, and scanned for use in this study.

### 10.2. Depositional Zone

The most pronounced change within Tahoma Creek has occurred between RKs 3 and 7, where debris flow deposition has killed large swaths of valleyfloor forests and re-routed Tahoma Creek (figure 20). In 1951, the stream was seen to flow through a narrow active channel, situated along the eastern margin of the valley. The forested area in the lower extents of this photo shows swaths of distinctive canopy textures, likely demarcating historical channel locations. By 1984, just prior to the recent suite of debris flows initiating in 1986, the valley showed significant forest mortality and channel widening, the result of debris flows occurring between 1967 and 1972. The damage is focused in the lower reach, and a narrow ribbon of bare gravel has become visible along the western margin of valley floor. By 1996, the passage of at least 14 debris flows removed extensive swaths of valley-floor forests, including a large percentage of forests in the upper reach. In the years since 1992, when debris flow activity slacked, sizable areas of this reach were recolonized by red alder, though many of these stands were again removed by a 2005 debris flows and the subsequent 2006 flood. Seen most recently in 2009, the extent of bare gravel within this reach has expanded significantly, and much of the forest in the upper extents of the photo is completely gone. In the lower extents of this reach, the channel has avulsed, now following the western margin of the valley within the recently cleared gravel swath.

#### 10.3. Fluvial Channel

Below RK 4.5, the channel is dominated by fluvial processes, sitting beyond the extent of most debris flow deposition. Below the marked avulsion between RK 3 and 4.5, the river course has changed little over the period of record. The 2006 flood did lead to a temporary avulsion in these lower reaches, with the majority of the flow jumping west into the woods near RK 0.8 (Paul Kennard, personal comm., 2012). The channel it incised is clearly visible in figure 23, and is locally over two meters deep. However, deeply rooted conifers growing on the floodplain surface prevented the channel from widening significantly; the new channel eventually choked with sediment near the original avulsion point, and flow returned to the main channel. Such events illustrate the lateral stability floodplain forests impart to these channels.



Figure 20: Repeat aerial photography, extending from RK 3 to 7. The Westside Road is visible along the western margin of the photos



Figure 21: Historical aerial photos showing channel avulsion at RK 2.

## 10.3.1. Channel Activity Near RK 2

Within the lower channel, the most notable re-routing over the past 60 vears has been a small avulsion near RK 2. This avulsion progressed slowly, already being evident in 1951 but remaining multi-threaded until at least 1989 (figure 21). The earliest photos of this site show a relatively clear channel to the west, though the main thread of flow sits to the east within several fresh-looking gravel ribbons. The western channel became largely vegetated by 1960, but was then reactivated sometime before 1969, and remained active until at least 1989. By 1996, the western path was again completely vegetated, while the eastern path had widened significantly to become the primary channel. This remained the case in 2012. Field examination of the western channel in 2011 showed shallow flow paths and some sand deposition, indicating only mild inundation during the 2006 flood. As of 2008, this abandoned channel sat roughly two meters higher than the adjacent active channel, indicating incision occurred sometime after 1989 (figure 22). Between a half-meter to a meter of this incision occurred between 2002 and 2008, presumably during the 2006 flood. Alder establishments on this, and similarly elevated surfaces to the east, occurred primarily in or soon after 1991, with an older cluster establishing in 1978. Both of these dates correspond to periods in the immediate aftermath of debris flows, and both follow relatively high peak flows recorded at the National gage, in 1977 and 1990 (figure 19).

Overall, this reach appears to undergo vertical change with a frequency

and scale not evidenced elsewhere in the lower basin. The combined evidence of the aerial photographs and alder ages suggest that aggradation has occurred between 1960 and 1969, and possibly sometime between 1978 and 1990, while incision occurred in 1977, 1990 and 2006. These dates make a compelling case that debris flows lead to transient aggradation in this zone, but that the channel re-incises relatively soon after debris flows cease, and particularly following the passage of a high flow.

A similarly elevated side channel, with an abandonment date in the years before 1929, sits just upstream of this recent avulsion (figure 22). This surface provides evidence that, at least locally, Tahoma Creek has aggraded in the past, possibly in response to debris flows occurring in the early decades of the 20th century as presented in section 11.

#### 10.3.2. Lower Channel

Below RK 2, isolated stands of conifers growing on low surfaces along the margins of the modern active channel provide abandonment dates for these surfaces. Two stands on opposing sides of the channel just upstream of the bridge both established in the years before 1948-50, while a low-lying forested island near RK 0.8 was last active sometime before 1910 (figure 23). Monthly precipitation values from Tacoma, WA indicate that 1909 experienced a notably wet November, making this year a likely candidate for a large flood. Similarly, two of the highest single-day precipitation records for Longmire, WA occurred in March, 1943 and December, 1946. Both are



Figure 22: Map showing channel near RK 2. The recently abandoned channel discussed in section 10.3.1 is shown along the black line, while the pre-1929 channel is shown with the red line. Individual trees with pith ages are indicated. Base map color ramp shows the locations height above the adjacent 2008 low-flow water level. The plot below shows the mean elevation of the two indicated abandoned channels in the bold lines, as compared to the modern active channel. Dashed lines represent the minimum and maximum elevations of their respective surfaces.



Figure 23: Map showing the age distribution of conifers cored on several young surfaces within the active channel. Even-aged stands are indicated by a single symbol, with the individual ages of the trees composing the stand noted. Base map color ramp indicates height above the adjacent low-flow water level, as of 2008

inferred to be rain-on-snow events, given warm daily temperatures (over 11 C for the 1943 event), and a loss of snowpack as measured at Longmire. These storms presumably led to significant local flooding, and both are plausible causes for the clearing of surfaces later recolonized by these stands.

### 10.3.3. Channel Width

The active channel of Tahoma Creek has widened noticeably since 1951, a common response to changing hydrology and/or sediment supply. To quan-

tify this change, aerial photographs from 1951 and 1984 were georeferenced to the 2009 imagery, and the lower four kilometers of the active channel was digitized for these three years. The mean active channel width, measured every 100m, has nearly doubled since 1951, with over half of this increase occurring between 1951 and 1984 (figure 24). However, the period from 1947 and 1955 was conspicuous for the lack of any intense precipitation events at Longmire, and generally mild flooding along the Nisqually. Close examination of the 1951 photos show more alder encroachment into the active channel. These observations, coupled with the fact that the widening from '51-'84 occurred in a relatively uniform manner along the length of the lower channel, suggest that 1951 may represent an unusually narrow and stable state for Tahoma Creek, and that channel widths in 1984 may be more in line with what was present during the hydrologically active decades of the 1920s and '30s. In contrast, widening over the later period was significantly more localized, and often reflected positions where the river actively attacked floodplain sidewalls. These side-sweeps created distinct scalloped cuts that were often filled with large logjams, formed as floodplain forests toppled into the channel. One such feature is nicely evident in figure 23 along the western sidewall near RK 0.5, producing the associated widening seen in figure 24. Aerial photos show that these localized incursions into the floodplain were caused primarily by the 2006 flood.



Figure 24: Mean channel width for indicated years, as measured from aerial photography. Mean values are shown as dashed lines.

#### 10.3.4. Change at the Bridge

The earliest useful records of Tahoma Creek are centered around the bridge. Highway 706 has crossed Tahoma Creek at the same location since 1909, with several photos from early bridge construction efforts providing a glimpse at the channel in the early years of the 20th century. Apart from the woody debris seen in the lower-left photo of figure 25, which is related to the recent construction of the pictured bridge, it is difficult to discern any major qualitative difference between the channels. The lack of fixed hard points in the photos make any judgement of the vertical position of the channel difficult. However, the channel in 1915 does appear to situated relatively high, with little relief between the bar and water surfaces. In 1918, following the 1917 flood, the channel appears to have incised into these sediments,



Figure 25: Repeat photography of the Tahoma Creek Bridge. Photos in 1915 and 1918 were both taken immediately after construction of the pictured bridges. The earlier concrete span was quickly washed out in a 1917 flood, when woody debris accumulated on the central pier and directed the flow against both channel walls. Log-stringer bridges, such as the one pictures in 1918, were in place until the modern structure was constructed in 1968.

particularly visible just upstream of the bridge. The 2011 photo presented here as a modern comparison represents the highest vertical position the channel as reached in recent years, and was dredged in the winter of 2012 to create just over two meters of freeboard below the bridge. Similar dredging efforts were noted by Walder and Driedger 1994a during the late '80s and early '90s, and have occurred more recently in 2006, 2008, 2010 and 2011. Dredging occurred at least several times between the late '90s and early '00s, though no records of exact dates were available.

# 10.4. Summary

Drawing a coherent line through the scattershot of analyses presented here poses something of a challenge. Perhaps the most conspicuous element is the lack of a clear response to the numerous debris flows over the past 40 years. While the channel has widened appreciably over this period, this is at least partially a function of the narrowness of the 1951 channel, which may not be representative of the mean state of Tahoma Creek, while widening over later years is likely more related to the 2006 flood than to any debris flow activity. The age of conifers growing on gravel surfaces provide evidence of energetic channel restructuring just before 1910, 1929 and 1948, all of which correlate well with possible flood events. The elevations of these surfaces suggest that Tahoma Creek has not changed its vertical position significantly over the past century. As noted above, the lack of significant channel change does not mean that sediment fluxes from Tahoma Creek have been consistent, but does suggest that these fluxes have not greatly exceeded the historical range of variability.

### 11. Dendrochronologic Reconstruction of Disturbances

Dendrochronologic methods have been shown to be effective at reconstructing past occurrences of geomorphic flows, including both floods and debris flows (Stoffel, 2008; Zielonka et al., 2008; Saez et al., 2011; Díez-Herrero et al., 2013). During such events, valley-floor forests are often inundated with sediment, resulting in impact injuries, burial, tilting or undercutting that are recorded as growth disturbances (GD) within the tree-ring record. These growth disturbances may variously manifest as cambial injuries, compression or tension wood, tangential rows of traumatic resin ducts (TRD), or abrupt suppressions of growth (figure 26). Conversely, survivors within a stand experiencing significant mortality may show growth releases as competition is removed.

The floodplain/terrace surfaces within the Tahoma Creek valley support a host of forests composed primarily of Douglas Fir (*Pseudotsuga menziesii*) and Western Hemlock (*Tsuga heterophylla*), intermixed with Pacific Silver Fir (*Abies amabilis*) and Western Red Cedar (*Thuja plicata*). Many of these stands have been inundated with sediment during recent floods and debris flows, and presumably have been subject to similar disturbances in the past. Reconstructions of fire activity around Mt. Rainier presented in Hemstrom and Franklin (1982) are available to distinguish these events from the sediment flows of interest, though both those records and field evidence do not indicate any major fires in the recent record. Broadly, the favorable position of these valley-floor forests reduces the potential for disturbances beyond these sediment flows to impact growth over much of the valley.

### 11.1. Methods

Tree-ring records from 158 conifers growing along the valley floor of Tahoma Creek were sampled during the summer months of 2011 and 2012. An increment borer was used to extract two cores from each tree. Sam-



Figure 26: Examples of growth disturbances within conifer tree-ring records used to reconstruct debris flow and flood events. Plates A and B show both TRDs and suppression following inferred debris flows. Plate C shows significant release following significant local stand mortality. Plate D shows a tree that has been buried by a debris flow from the mid-19th century, but survived and continue to grow above the burial line. A recent debris flow, likely 2006, finally killed this tree. Plate E shows mortality from burial and injury occurring near RK 8.

pling occurred between RKs 0 and 8, with a focus between RKs 3 and 5, covering the depositional zone of debris flows and the fluvial reaches below. Specific sites were selected on the basis of their potential to record disturbance events, as evidenced by either the presence of modern sedimentation or through a combination of age structure, surface characteristics and valley position that suggested historical disturbance. Once collected, these cores were dried, mounted and sanded, after which the rings counted and ring-widths measured. Both counts and measurements were obtained using WinDendro, a semi-automated image analysis program, after making high-resolution scans of all cores. For *P. menziesii*, dating errors were checked through comparison with a previously developed chronology for Tahoma Creek (Brubaker

et al., 1992), while T. heterophylla were dated on the basis of several marker years identified within the basin. Consistent marker rings were relatively sparse for the years before 1860, and so T. heterophylla chronologies before this date are subject to some uncertainty. The few A. ambilis and T. plicata cores were cross-dated between themselves, but the age control is relatively poor for these cores, and so were used primarily for age of establishment or the corroboration of events recorded in other species.

After counts and measurements were performed, each core was examined for the growth disturbances noted above, and the year and type of each disturbance was charted. These records were compiled and years with a number of disturbances significantly above background levels were noted. Given some uncertainty in the age control for older T. heterophylla, and the modern trend of debris flows to cluster, sequential years of increased GD rates were sometimes combined to describe a single "event." For example, the 1697 event is defined by GDs in 1697, 1698 and 1699, with three, two and one GDs respectively. In contrast, the background rate of GDs in the surrounding decades was 0.18 per year. Once such possible events were identified, the trees defining the event were plotted spatially and both the number of disturbances and the logical consistency of their spatial distribution was examined before a decision was made on whether or not to accredit them to a geomorphic event. Those that were deemed to be the result of a geomorphic flow were described as either high- or low-confidence, again based on the number and location of the disturbances. These events were then classified as either floods, debris flows, or both, based on the percentage of disturbances occurring high and low in the basin (figure 27). Beyond these three classifications, a possible date for the Tahoma Lahar and landslide activity from the western valley wall near RK 3.4 were also identified.

## 11.2. Results

In total, 17 debris flows and 18 floods from between c. 1500 AD to present were identified in the dendrochronologic record, comprising 29 unique events. Of these, 20 were classified as high-confidence (table 2). An examination of the distribution of pith ages for the cored trees also showed two distinct periods of establishment.

When compared to the modern records of debris flows, this chronology only identified three of the 27 known events. The relative sparsity of the reconstructed record is attributable to both the loss of information that occurred as valley-bottom forests were cleared during prior debris flows, and the fact that dendrochronologic methods can only record events that cause significant and widespread damage.

The most notable feature of the chronology, outside of the modern events, is a suite of debris flows occurring between 1840 and 1900, with a particularly high concentration between 1840 and 1860. These early events were followed by a period of significant recolonization between 1870 and 1890 - 25% of the 158 trees cored reaching coring height during these years (figure 28). The spatial distribution of these establishments closely mirrors the valley areas

Year	Event Type	disturbances recorded	Sample depth	Percent affected	Establishments
1508	Tahoma Lahar?	3	9	33%	na
1530-1563	Establishment	na	na	na	12
1611	Debris flow	2	33	6%	na
1643	Landslide	-	38	11%	na
1649	Debris flow	5	38	13%	na
1685	Flood	4	44	9%	na
1697	Debris flow	6	45	13%	na
1730	Flood	5	51	10%	na
1753	Debris flow	5	55	9%	na
1791	Flood	4	66	6%	na
1826	Flood	7	73	10%	na
1820	Debris flow	4	73	5%	na
1840	Debris flow	5	81	6%	na
1847	Debris flow flood	7	84	8%	na
1853	Debris flow	7	87	8%	na
1855	Flood debris flow?	7	87	8%	na
1877	Flood	4	98	4%	na
1880	Debris flow	7	101	7%	na
1890	Flood	4	120	3%	na
1895	Debris flow, flood	12	129	9%	na
1870-1896	Establishment	na	na	na	38
1905-8	Floods, debris flows?	11	135	8%	na
1911-12	Debris flow, flood	5	136	4%	na
1925-6	Flood	11	144	8%	na
1936	Flood	9	148	6%	na
1959-60	Flood, debris flow?	16	155	10%	na
1966-1968	Outburst floods?	16	158	10%	na
1988	Flood, debris flow?	8	158	5%	na
1991	Debris flow	20	158	13%	na
1993-1995	Debris flows	19	158	12%	na
2007	Flood, debris flow	14	151	9%	na
2008-2009	Floods	13	151	9%	na

Table 2: Table of events reconstructed from dendrochronologic analysis. Bolded entries are high confidence, italicized entries are low confidence. Floods and debris flows were differentiated by spatial pattern of disturbances.

that have experienced disturbance in the recent decades, and in particular, follow the western course of the modern channel (figure 27). There is no indication whether this period of recolonization was the result of a decrease in debris flow frequency and/or intensity, or simply an avulsion that moved the river to its eastern course, though the extent of recolonization occurring above and below the likely avulsion suggest that at least the former occurred. There is a reasonable overlap between the locations of these late-19th century stands and the swaths of distinctive canopy textures seen in the 1951 aerial photographs, suggesting that the latter are indications of the passage of the former.

Conifer establishments begin increasing in 1820, pre-dating the debris flows mentioned above by several decades (figure 28). The actual passage of these debris flows manifests in the low number of establishments occurring between 1855 and 1865, with the lag between these dates and those of the debris flows agreeing well with eccesis intervals for douglas fir in such sites (Pierson, 2007). The majority of these earlier establishments occur lower in the basin, including several shown in figures 22 and 23. While diffuse, this increase in establishments may be related to low-confidence debris flows dated to 1831 and 1840. However, the lack of similar establishments following the later, and likely more intense, debris flows suggests that these trees may have colonized surfaces cleared by flooding. A probable candidate is a flood dated to 1826. These trees may also be related to a fire dated to 1820 (Hemstrom and Franklin, 1982), though the authors mentions that this fire



Figure 27: Map showing recorded disturbances defining several high-confidence events. Maps extend between RK 0 and RK 8. Red triangles represent trees showing disturbance in a given year, green triangles represent trees that established during the indicated period. White circles represent trees with a ring record covering the indicated year.



Figure 28: Distribution of pith ages for cored conifers, clustered in 10-year increments and centered on decades. The dates of debris flow events reconstructed through growth-disturbance analyses are indicated, with high-confidence events filled black and low-confidence events filled white.

did not generally burn down onto alluvial surfaces.

Prior to 1800 AD, only four debris flows were identified, of which two were considered high-confidence events. The relative sparsity of the record in this early period is at least partially a function of the loss of information that occurred when the debris flow sequences of both the mid-19th and late-20th centuries caused significant mortality in valley bottom forests. The pulse of establishments occurring between 1530 and 1560 likely represents the recolonization of the surfaces cleared by the Tahoma Lahar, which has previously been dated to sometime between 1480 and 1535 AD in previous work (Crandell, 1971) and refined here to 1508 AD based on a number of tree-ring records covering these years. The smaller pulse in the last decades of the 15th century may be also be delayed colonizations on these surfaces, though this is uncertain.

### 12. Discussion - Climate and Debris Flows

To assess whether temperature, precipitation or glacial mass balance may be first order controls on debris flow frequency, we compared our chronology with several historical reconstructions of these properties. While temperature and precipitation ultimately drive glacial mass balance changes, the complex timing of the glacial response to both climate change and interannual variability over decadal time-scales creates asynchrony between the driving forces and resulting glacial response (Roe and O'Neal, 2005; Roe, 2011) This asynchrony allows us to examine glacial mass balance as something of an independent variable.

Mean annual temperatures were obtained from a dendrochronologic record centered at Longmire, WA (Graumlich and Brubaker, 1986), while periods of glacial retreat were documented through the dating of moraines, using dendro- or lichenometric methods (Sigafoos and Hendricks, 1972; Burbank, 1981). We use a 500-year dendrochronologic reconstruction of the PDO (MacDonald and Case, 2005) as an indicator of the regional hydroclimatology, as both local glacial dynamics and river discharge have been shown to follow this index (Cayan et al., 1998; Czuba et al., 2012a).

The resulting comparisons between our debris flow/flood record and the PDO, mean annual temperature and moraine ages are shown in figures 29 and 30. The PDO shows no clear relationship with either flooding or debris flows, as both events occur with equal frequency during warm, cool and neutral PDO periods. This lack of correlation does not necessarily imply

that climatically driven variations in precipitation are inconsequential for debris flow frequency. Most simply, intense rainfall events should be the most common triggers of debris flows - records of daily precipitation from Longmire, dating back to 1909, show no correlation between such events and the PDO, suggesting that this index may not be an appropriate proxy for our study.

Temperature does not show any direct relationship with flooding, nor do debris flows appear to be more prevalent during warm or cool periods. There is some association between debris flows and turning points in the temperature record, particularly transitions from cooling to warming trends. These periods of warming would logically correlate with the onset of glacial retreat, though the moraine record shows that this is not always the case. While many of the documented periods of retreat do sit nicely near the onset of warming trends (e.g. 1620s, 1750s, 1903), there are a number of moraines that formed during warm peaks or cooling trends (e.g. 1660s, 1690s, 1820s).

Ultimately, it is periods of glacial retreat that best correlate with our dated debris flows. Moraines from the South Tahoma Glacier dated to 1843 and 1864 nicely match high-confidence debris flows dated to 1847, 1853, 1880 and 1895, and nearly every debris flow recorded before the 1930s sits at or near a dated period of retreat. Of the five such periods of retreat documented between 1600 and 1800, four are matched by contemporaneous debris flows, and no debris flows were documented during the intervening periods. Given the extremely rapid retreat over the past decades, the modern debris flows corroborate this connection. While periods of warming within the temperature records do show decent correlation with both debris flow events and moraine ages, debris flows dated to 1623, 1697, and 1880 all occur during years of constant or cooling temperatures, yet match dated periods of retreat well. This suggests that it is the retreat that is the first order control, and not temperature. Similar results have been observed in Kautz Creek, where a debris flow chronology based on lichenometric, dendrochronologic and C14 dating showed that events generally occurred during warming periods in the decades following the onset of glacial retreat (Legg, unpublished data).



Figure 29: PDO index over the past 500 years. Before 1950, values are dendrochronologic reconstructs from Macdonald and Case. After 1950, values are based on observationally data as presented by Mantua. High- and low-confidence debris flows and floods are indicated by black- and white-filed symbols, respectively. Event y-values are given as the PDO index of that year.


Figure 30: Mean annual temperature (11-year centered mean) compared with debris flow and flood events from this study. Temperature before 1950 is taken from a dendrochronological analysis centered at Longmire, WA, with records after this come directly from the station at Longmire. Debris flow and flood events are plotted with y-values corresponding to the running-mean temperature of the year of the event. Periods of mountain-wide glacial retreat are shown as grey vertical bars (Sigafoos and Hendricks, 1972; Burbank, 1981), with moraines dated for South Tahoma Glacier shown as solid black lines (Sigafoos and Hendricks, 1972)

#### 12.1. Modern vs. Historical Debris Flows

The modern debris flows have significantly reshaped the channel between RKs 4 and 7. While impressive, the reconstruction here suggests that similar events have occurred in the recent past. Given the extent of disturbances and the significant reconolonization that occurred following the 1847-1853 debris flows, this sequence of debris flows was likely as destructive as the modern suite. For events prior to 1847, no major periods of recolonization were noted, nor do the relatively sparse record of disturbances give much indication of their extent. However, between the inferred mortality of the 19th century events and the readily observed modern mortality, much of the dendrochronologic records of debris flows, and particularly records of smaller events, have likely been destroyed. That debris flows prior to 1840 have remained in the record suggests that they were similar in scope to the more recent events.

However, there are indications that the modern events are novel. Conifer stands near RK 8 were largely established between 1530 and 1560, presumably colonizing bare surfaces created by the 1508 Tahoma Lahar. In the following centuries, trees within these stands showed remarkably complacent growth, indicating little disturbance from the passage of subsequent debris flows. However, by 2012, the majority of these trees were dead, killed by impacts from boulders and burial that frequently exceeded several meters. While these trees did record some disturbances during the 1986-1992 debris flows, much of this mortality appears to have been related to the 2006 flood and associated debris flows. The unprecedented damage from the 2006 event could reflect an increase in the peak intensity of flood events, an increase in the size of debris flows owing to enhanced sediment availability, or both.

## 12.2. Outburst Floods as Triggers of Debris Flows

Many of the modern debris flows in Tahoma Creek have been attributed to outburst floods from South Tahoma Glacier (Walder and Driedger, 1994a). These outburst floods, in turn, are thought to be the result of a rapid increase in the water pressure at the glacial bed, which causes a sub-glacial network of cavities to expand, link and catastrophically drain Driedger and Fountain (1989); Walder and Fountain (1997). These increases in fluid pressure can be driven by both rapid melt or heavy rainfall, and both have been associated with debris flow events around Mt. Rainier. The actual evidence for outburst floods from South Tahoma Glacier is largely circumstantial, as none have been directly observed. However, outburst floods elsewhere on the mountain are well-documented (Hodge, 1974; Driedger and Fountain, 1989), and they remain a plausible source for the large volumes of water incorporated within dry-weather debris flows. This connection is supported by the observation that all dry-weather debris flows occurred between 2pm and 9pm, matching the timing of the diurnal peak in streamflow. Regarding wet-weather debris flows, a recent outburst flood from the Nisqually Glacier was recorded at a stream gage sitting at Longmire, showing a spike in river stage that peaked at over 250% of the background flow (Scott Beason, personal comm. 2012). This spike lasted only several hours, and was preceded by the passage of a moderately-sized rainstorm. This event was small, but provides solid documentation that periods of rain may also trigger outburst floods.

Given the extremely rapid increase in streamflow associated with outburst floods, the case for these events being common triggers of debris flows is compelling. However, there is, at present, no strong evidence to suggest whether or not outburst floods are more prevalent during periods of glacial retreat. As such, outburst floods alone do not provide a strong physical explanation for the observed correlation between such periods and debris flows. Another possible explanation for this connection would be the increased sediment availability that accompanies glacial retreat. At first glance, such an explanation does not seem suitable for the 1847-53 debris flows, given that a moraine dated to 1864 sits only a few tens of meters back from the neoglacial maximum extent of South Tahoma Glacier dated to 1843. However, reconstructions of the glacial mass balance shown in Burbank 1982 indicate that, while the retreat of glaciers was most rapid from 1910-1940, a significant percentage of the mass loss occurred from 1850-1890 but resulting in thinning of the glaciers, with little associated retreat.<sup>5</sup> This thinning would de-buttress lateral moraines, presenting the opportunity for debris flows to initiate and flow along glacial margins. Sedimentological evidence points to

<sup>&</sup>lt;sup>5</sup>1850 simply represents the earliest date for which consistent climatic data was available for study, and not necessarily a transition point in the glacial record

such debris flows occurring along the sides of the Carbon Glacier, which has not retreated over the past century (Nylen 2004; Paul Kennard, personal comm. 2012). Given that outburst floods have been observed to issue midglacier and incise into lower stagnant ice on both the South Tahoma and Nisqually Glaciers (Driedger and Fountain, 1989), these two explanations are not mutually exclusive.

Both the recent debris flows since the late 1960s and the suite of debris flows occurring in the mid-19th century have been concentrated in pulses of high activity in the decades following retreat. In contrast, the debris flows reconstructed before 1800 are all individual events that occur at nearly the exact same time as retreat events. There are several factors that may explain this mild paradox. First, as all moraine ages are given as the age of the oldest tree or lichen found its surface, all dates represent upper limits. Between potentially lengthy eccesis times and the inherent difficulty in locating the oldest individual present on a surface, it is likely that the actual retreat occurred somewhat earlier than indicated, particularly so for older moraines. This would provide an explanation for why the two earliest dated debris flows appear to slightly pre-date their associated periods of retreat. Additionally, sediment availability would presumably be higher for the debris flows occurring earlier in a sequence, making them larger and more destructive and so more likely to be recorded in the tree-ring record. That the largest debris flows of both the two modern sequences occurred during the first two years of activity provide support for this idea. Lastly, it is entirely plausible that the overlap of debris flows and retreat in the years before 1840 is a fluke, given only four such events were documented. Even so, the strong signal of debris flows following the LIA retreat of the 1840s remains conspicuous.

#### 12.3. Correlation with Re-advancements

Both recent periods of debris flows have occurred during glacial standstills or re-advancements situated within the overall trend of retreat (figure 31). While similarly resolved measurements of the South Tahoma Glacier terminus are not available for earlier periods, there exists a reconstruction of glacial mass balance for Mt. Rainier based on measured temperature and precipitation records extending back to 1850 (Burbank, 1982). Within this reconstruction, transitions from negative to positive mass balance occurred in 1878, 1891, 1907, 1915 and 1932. Debris flows are inferred to have occurred in Tahoma Creek in 1880, 1895, 1905-1908, and 1912, though only the earliest two are considered high confidence events. The close agreement between these dates is intriguing - if debris flows are particularly likely to occur during slight re-advances or standstills situated within overall retreat, then inter-annual climatic variability may be as much a factor in determining debris flow frequency as the mean climatic state. Hodge (1974) noted the occurrence of outburst floods issuing from the Nisqually Glacier in 1968, 1970 and 1972, a period when glaciers along the southern flank of Mt. Rainier were advancing, providing some suggestion that these events may correlate with periods of advance. We also note the overlap of these dates with the



Figure 31: Terminus location of South Tahoma Glacier, as inferred from aerial photographs, topographic maps and dated moraines. For aerial photos, the terminus was simply noted as the position of the outlet stream, and so does not differentiate between active and stagnant ice. The large discrepancy between the indicated terminus position from Nylen (2004) for the period around 1960 is likely a function of this distinction. An alternate interpretation of a 1958 aerial photograph, in which the terminus is noted as a fresh, wet face of ice higher in the basin, produces a much better match. The two recent periods of high debris flow frequency are noted as vertical grey bars.

'67-'72 sequence of debris flows within Tahoma Creek. However, while these observations present an interesting possibility, there is currently no obvious physical explanation for this relationship, nor, to our knowledge, has such a correlation been suggested or observed before. The fragments of historical records we piece together here do not provide sufficient evidence to make a solid assertion, but do point to a line of inquiry for future research.

# 12.4. Local and Regional Comparisons

While Tahoma Creek has stood out in recent decades for the frequency of its debris flows, such events are common to pro-glacial drainages throughout the Cascades. To assess if there is coherency between the timing of debris flow events, both locally around Mt. Rainier and more broadly through the region, we compare the chronology for Tahoma Creek to existing records. Reconstructions of debris flow events prior to 1900 exist for Kautz Creek (Nick Legg, personal comm., 2012), and 10 drainages around Mt. Shasta (Hupp et al., 1987). Over the last century, data exists for several additional drainages around Mt. Rainier, though records before 1980 are exclusively for the Nisqually River. A chronology of debris flows on Mt. Hood was obtained from Tom Degroo (personal comm., 2012). All of these records are necessarily incomplete to various and unknown degrees, making detailed interpretation tenuous. Treating outburst floods on the Nisqually as part of a common pattern<sup>6</sup>, the mid-'80s and years surrounding 1970 appear particularly active.

 $<sup>^{6}</sup>$ The Nisqually Glacier fore-field is similar to that of the South Tahoma Glacier, and so the documented occurrence of outburst floods that did not transform into debris flows is



Figure 32: Debris flows and outburst floods for various basins on Mt. Rainier (this study; Nick Legg, personal comm., 2012; Driedger and Fountain 1989), as well as compiled records for Mt. Hood (Tom Degroo, personal comm., 2012) and Mt. Shasta (Hupp et al., 1987). For the pre-1900 records on Mt. Shasta, periods of increased activity as mentioned in the original publication are plotted. Note the vertical scale change at 1900.

There is some sense of increased activity in the late '20s and '30s, though no debris flows from this period were documented in either Tahoma or Kautz Creek. This period was particularly active on Mt. Shasta, were a series of outburst flood-generated debris flows caused significant damage in 1924-26 (Hill and Egenhoff, 1976). These events occurred during a period of exceptionally fast local glacial retreat, presenting some evidence that the connection between retreat and debris flows seen in Tahoma Creek is not a purely local one.

Before 1900, common periods of recorded debris flow activity include 1650-1700 and 1840-1870. The earlier of these periods corresponds to the warmest decades outside of the modern era (Graumlich and Brubaker 1986; figure 30), precipitating retreat in many of Mt. Rainier's glaciers. The later period marks the end of the LIA and the onset of modern retreat. Both such periods provide some additional weight to the connection between glacial mass balance and debris flow frequency presented here. Within the modern records, debris flows from all sources show a tendency to cluster over periods of five to ten years. This may be partially a function of the intermittency of detailed record keeping, though similar tendencies are seen within the Mt. Shasta chronology, which is based entirely on dendrogeomorphic methods.

Regardless of completeness, the modern records clearly demonstrate that

surprising. Walder and Driedger speculate that the relative infrequency of these outburst floods prevent the formation of a deeply incised gully in the pro-glacial sediments, such as is seen in front of the South Tahoma Glacier. The lack of immediate access to such steep sidewalls may prevent floods from bulking appreciably.

debris flows are a common occurrence with the Cascade Range, and with the possible exception of Mt. Hood, there is no clear indication that their frequency has increased in recent decades.

### 13. Synthesis of Results

Taken as a whole, the three sections above permit some broad statements to be made about sediment fluxes within Tahoma Creek. The debris flows of the past few decades appear to be part of a larger pattern of similar events that have occurred relatively frequently over the last 500 years, and most notably, during the end of the LIA. While these debris flows do appear to translate into increased channel mobility, and presumably increased sediment fluxes, within the lower basin, the recent events do not appear to have perturbed the system beyond historical limits of variability. Aggradation within the lower basin has been minimal and localized, and does not appear to represent a transition to a generally aggradational state. Looking back over the past 50-100 years, there is no clear indication that the channel is functioning differently now than at any other point within this period. While the width of the active channel has increased significantly since 1951, much of this increase manifested as the removal of active channel vegetation or the specific impact of the 2006 flood. Beyond this singular event, the evidences suggests that widening has occurred primarily within the confines of the historical active channel, and may reflect the quiescence of the hydrology in the years before 1951 more than the intensity of years following.

In contrast to the lower basin, the upper reaches have been dramatically altered by the recent sediment loading, and the wide-spread mortality of valley-bottom forests that have remained unaffected by previous events does give some sense that recent debris flows have been either more persistent or more intense than anything seen since the passage of the Tahoma Lahar in the early 16th century. Whether this reflects a change in sediment availability or local hydrology is unknown.

Our reconstruction of debris flows within Tahoma Creek presents several intriguing possibilities in regards to the interplay between glacial mass balance, outburst floods and debris flows. Debris flows within Tahoma Creek appear to be closely associated with periods of glacial retreat, though there is some evidence that short periods of positive mass balance punctuating retreat may be the most direct triggers. Purely on the basis of sediment availability, this connection is reasonable, and regional debris flow chronologies show some support for this idea. The role of outburst floods, and how their frequency may change as a function of climate, remains unclear.

## 14. Conclusion

Debris flows are a common occurrence within pro-glacial basins throughout the Cascades. The relative frequency of these events roughly correlate with changes in glacial mass balance, but the complex interactions between climate means, climate extremes and glacial responses that ultimately produce this correlation remain difficult to disentangle. While it is reasonable to assume that recent climate change has the potential to alter hydrologic and geomorphic systems around the Pacific Northwest, the morphology of Tahoma Creek has not borne witness to any such change. However, recent debris flows do appear to drive accelerated channel mobility, and so presumably increased sediment fluxes, within the lower basin. While the history and morphology of Tahoma Creek may allow it to accommodate such increased loads with only minimal geomorphic response, aggradation within low-angle reaches downstream remains possible.

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