



Vanishing volcanics: Implications of mass wasting deposit sediment heterogeneity on downstream abrasion and the fluvial hazard cascade

Brian C. Pinke ¹, Allison M. Pfeiffer ^{*1}, Jonathan A. Czuba ², Finn Coffin ¹

¹Geology Department, Western Washington University, Bellingham, USA, ²Department of Biological Systems Engineering, Virginia Tech, Blacksburg, Virginia, USA

Author contributions: *Research conceptualization:* B. P., A. P., J. C. *Methodology:* B. P., A. P. *Formal analysis:* B. P., A. P. *Writing:* B. P., A. P. *Figure drafting:* B. P., A. P. *Data curation:* B. P., A. P. *Investigation:* B. P., F. C., A. P. *Funding acquisition:* A. P., J. C., B. P. *Project administration:* B. P., A. P., J. C. *Supervision:* B. P., A. P., J. C. *Editing:* B. P., A. P., J. C. *Review & editing:* B. P., A. P., J. C., F. C.

Abstract In the Pacific Northwest, over half of all catastrophic mass wasting events occur on volcanic slopes, making volcanic edifices significant contributors of sediment to downstream channels. The grain size distribution (GSD) and clast abrasion rates of mass wasting deposit sediment can strongly control the downstream fate of these large sediment pulses, mediating the magnitude and timing of the fluvial hazard cascade. Here, we build on prior research from Glacier Peak, quantifying the GSD, clast strength, and simulated downstream sediment abrasion of four additional volcanic mass wasting deposits: one on Mt. Meager and Mt. Adams, and two on Mt. Rainier. Deposits vary substantially in GSD ($D_{84}/D_{50} = 3.8\text{--}8.0$) and include diverse volcanic rock types. We find that vesicularity and hydrothermal alteration, which weaken volcanic rocks, strongly influence downstream abrasion. Individual clast abrasion rates span four orders of magnitude, with simulated volumetric losses up to 86% over 100 km for the weakest deposit. Comparison to a global dataset indicates these rocks are of typical volcanic strength. Importantly, sedimentary rocks tend to be substantially weaker, suggesting sedimentary pulses could experience even greater attenuation. Our findings highlight the role of bed material abrasion in coarse sediment budgets, with implications for source-to-sink hazards in volcanic and other river systems.

Non-technical summary Large landslides are common on the volcanoes of the Pacific Northwest. These events can dump enormous volumes of sediment (ranging from silt to boulders) into rivers. That sediment has the potential to travel downstream and exacerbate flooding. This study examines five large landslide deposits at four volcanoes to understand what happens to these rocks as they travel downstream. We measured rock size, strength, and type in the field, then used computer models to simulate how the material would change after 100 km of river transport. Our results show that these volcanic landslide sediments vary substantially in strength and size, both within and between sites. The weakest rocks break apart rapidly during river transport: at the weakest deposit, ca. 86% of the original rock volume would be ground into fine particles within 100 km. Only the strongest rocks survive long downstream journeys. This research has implications for flood risk management. Communities may overestimate how much sediment will reach populated areas, since much of it may wear away during transport. As climate change increases volcanic landslide frequency through melting glaciers and intense rainfall, understanding this natural “filtering” process helps us better predict downstream hazards and improve emergency planning.

1 Introduction

With more than half of all catastrophic mass wasting events in the Pacific Northwest of North America occurring on slopes of volcanic rocks and regolith, these volcanic landscapes disproportionately supply large volumes of sediment to river valleys (Geertsema et al., 2006; Sterling and Slaymaker, 2007). Efforts to study and understand mass wasting vary from studying deposits in the field and using remote sensing techniques such as LiDAR and InSAR (e.g. Sohn et al., 1999; Major et al., 2005; Roberti et al., 2017; LaHusen et al., 2020; Handwerger et al., 2022) to modeling (Hutter et al., 1994;

Rickenmann et al., 2006; Schraml et al., 2015), and creating scaled flume experiments (Iverson, 1997). These studies largely focus on initiation, runout, and, importantly, direct hazards caused by a given event (Davies et al., 2013). Comparatively few studies consider what happens to these large mass wasting deposits after deposition into river valleys; for example, how the initial characteristics of mass wasting sediment affect how it subsequently evolves as it is transported through the fluvial system (Sutherland et al., 2002; Roda-Boluda et al., 2018).

The influx of sediment from these catastrophic mass wasting events can propagate downstream as cascading hazards (Yanites et al., 2025), potentially aggrading channels and thereby exacerbating flood hazards

Received:
October 3, 2025
Accepted:
April 7, 2026
Published:
July 28, 2026

Editor-in-Chief:
Alice Lefebvre
Associate Editors:
Jackie Cockburn
Serina Diniega
Managing Editor:
Olivier Gourgue
Copy Editor:
Ciro Cerrone
Production Editor:
Ciro Cerrone

Reviewers:
Karen Gran
James Christie

The work is licensed
under CC BY 4.0
International.

*Corresponding author: pfeiffa@wwu.edu

(Slater et al., 2015) and posing risks to life and property. Given that mass wasting events are expected to occur more frequently in the future due to climate change, e.g., via increased rates of glacial melt/retreat, accelerated freeze-thaw cycles, and increased rainfall intensity (Guzzetti et al., 2007; Stoffel and Huggel, 2012), understanding the variable strength of mass wasting deposit sediment and resultant downstream evolution of these sediment pulses is important. Past research on sediment pulse morphodynamics has largely relied on hydraulic flume studies (Cui et al., 2003; Sklar et al., 2009; Humphries et al., 2012) or simple numerical models that emphasize pulse volume and grain size distribution (GSD), but has lacked constraints tied to the underlying source material characteristics (An et al., 2017). The GSD is a major control on the downstream response (Sklar et al., 2009; Ahammad et al., 2021): finer grains will pass through the system quickly, while coarser grains will move more slowly, and mixed grain size distributions will have different impacts on river bed mobility than uniform distributions (Ahammad et al., 2021). Furthermore, the GSD of sediment in a river upstream can impact the rate of river incision and the river slope downstream (Sklar, 2024). This highlights a key question: what is the GSD of these volcanic mass wasting deposits that are both large in magnitude and common in the Pacific Northwest?

The initial grain size of sediment supplied from hillslopes depends on a range of physical and chemical mechanisms. Previous work demonstrates that at higher elevations, physical weathering is dominant, leading to coarser clasts, and at lower elevations, chemical weathering is dominant, leading to finer clasts (Sklar et al., 2017; Roda-Boluda et al., 2018). In addition to climate, lithology is a key factor in controlling sediment size (Sklar et al., 2017; Roda-Boluda et al., 2018). Generally, the stronger the intact bedrock is, the larger the resultant grains are due to fewer fractures and greater spacing of discontinuities. Grains with higher intact rock strength are less likely to break down through physical weathering. In crystalline rocks, the type and size of mineral grains influence the resulting particle sizes produced by chemical weathering, with coarser minerals generating coarser particles (Sklar et al., 2017; Roda-Boluda et al., 2018). Using a rock Schmidt Hammer as a proxy for strength, Roda-Boluda et al. (2018) demonstrated that fresher and less fractured bedrock is typically stronger and that these stronger rocks tend to weather to coarser clasts, though there may be a large range of grain sizes derived from one lithologic group. These previous studies suggest that, in the case of mass wasting deposits, fresh rock at high altitude would potentially weather into larger clasts due to physical weathering dominance. However, unique to volcanic terrain, hydrothermal alteration generally physically and chemically weakens rock (Watters et al., 2000; del Potro and Hürlimann, 2009; Schaefer et al., 2023). As a result, volcanic mass wasting material sourced at high altitude could potentially provide smaller grains to the landscape than otherwise expected. Further, the weakening of rock mass associated with hydrothermal alteration can systemically weaken volcanic slopes, mak-

ing volcanic edifices particularly prone to mass wasting (Watters et al., 2000; John et al., 2008; Finn et al., 2001) and thus readily delivering sediment down valley. There is a lack of research on sediment grain size delivered specifically from volcanic mass wasting events. This is important because grain size and the implicit factors that control it also play key roles in the evolution of clast size after the sediment enters the river channel.

Once sediment is in the channel, downstream fining occurs via selective transport in which smaller and lighter grains are carried farther, and via abrasion, as clasts collide and break down to smaller grains (Sklar, 2024). Abrasion is the combination of attrition, the gradual wear of the surface into fine particles, and fragmentation, the loss of larger particles (Attal and Lavé, 2009; Kuenen, 1956). Similar to its control over GSDs delivered from hillslopes, lithologic rock strength is an important factor controlling abrasion rates, with abrasion rates varying by several orders of magnitude between strong and weak lithologies (O'Connor et al., 2014; Pfeiffer et al., 2022). Experimental work has shown that at low particle velocities, abrasion rates depend largely on lithologic strength; however, at higher velocities, abrasion rates can increase (Attal and Lavé, 2009). As such, the strongest rock types supplied to the river bed will largely dominate the bedload of the river (Attal and Lavé, 2009). Lithologic strength is correlated with density such that stronger rocks are typically denser and weaker rocks are less dense (Pfeiffer et al., 2022). In general, denser rocks will preferentially deposit upstream, while lighter rocks will transport farther downstream (Viparelli et al., 2015). Combining these effects, the downstream transport of mass wasting deposit sediments should be controlled by the source material lithology via its control on grain size, strength, and density distributions.

Despite decades of research on mass wasting event mechanics and hazards, we lack a strong predictive framework for how resultant sediment pulses transport downstream. Here, we aim to improve source-to-sink models of sediment fluxes by asking: what is the variability in the sediment size, strength, and density of volcanic mass wasting deposits, and how do those variables translate to downstream coarse sediment impacts? Pfeiffer et al. (2022) constrained the GSD, lithology, and abrasion variability of a volcanic mass wasting deposit along the Suiattle River at Glacier Peak, and demonstrated that much of the coarse material abrades to fines in the first 10-20 km of transport. We build on this work by quantifying clast strength and grain size distribution of four additional volcanic mass wasting deposits in the Pacific Northwest: along the Lillooet River at Mt. Meager, White River at Little Tahoma Peak (Mt. Rainier), Kautz Creek at Mt. Rainier, and Salt Creek at Mt. Adams, to test the hypothesis that sediment characteristics of volcanic deposits vary substantially between sites, which in turn has significant implications for the downstream effects of those events. We use the collected data, along with a downstream abrasion model to test the influence of rock strength on abrasion and transport and compare strengths to a global compilation.

2 Geologic and Geomorphic Setting

This study focuses on five field sites (Figs. 1 and 2, Table 1) in the Pacific Northwest of Canada and the USA: Lillooet River at Mt. Meager, Suiattle River at Glacier Peak, White River at Little Tahoma (Mt. Rainier), Kautz Creek at Mt. Rainier, and Salt Creek at Mt. Adams. The Canadian Coast Range and North Cascade Range are dominated by alpine environments and braided gravel-bedded rivers. While the bedrock lithology of the region is highly varied (Schuster, 2005; Erdmer and Cui, 2009), volcanic terrain contributes a high percentage of mass wasting sediment to the fluvial systems (Geertsema et al., 2006). Similarly, although the sediment sources of these deposits are all stratovolcanoes, the volcanoes vary in lithology, degree of hydrothermal alteration, and glacial erosional histories. The deposits also result from a variety of mass wasting processes, ranging from hydrothermally altered slope collapse to debris flows triggered by glacial outbursts. These differences make the five selected sites well-suited for comparing, contrasting, and quantifying the degree of variability in mass wasting deposit characteristics with a focus on fluvial processes and downstream transport. Throughout the text, we refer to each deposit/event by the name of the mountain, except for the two Rainier deposits which we refer to by source zone (Kautz Creek and Little Tahoma).

2.1 Mt. Meager (Lillooet River)

Situated approximately 150 km north of Vancouver, British Columbia, Canada, Mt. Meager (2,650 m) is generally considered the northernmost edge of the Cascade volcanoes and lies within the Canadian Coast Range (Figs. 1 and 2a). The flanks of the volcano are composed of volcanic rocks that are Pleistocene through the Holocene in age, with the oldest recorded eruption approximately 2.2 Ma ago and the youngest just 2,350 years ago (Guthrie et al., 2012). Eruptive units consist of porphyritic rhyodacite and extrusive tuff, breccia of granite and aphanitic inclusions, and dacitic flows (Read, 1990; Roberti et al., 2017). Active hot springs and bedrock water seeps around the mountain and have resulted in hydrothermally altered volcanic rock that is highly fractured and weakened (Guthrie et al., 2012; Roberti et al., 2017). Freeze-thaw cycles, along with glaciation and snowmelt, result in changing bedrock and soil-water pore pressure that contribute to mass wasting events from the edifice, making Mt. Meager one of the most landslide-prone regions in Canada (Guthrie et al., 2012).

The August 6, 2010, rockslide-debris flow occurred on the south flank of Mt. Meager above a small glacier at the head of Capricorn Creek valley, which feeds into the Lillooet River (Fig. 2a). The slide was likely caused by changes in groundwater pore pressure, resulting from the end of summer snowmelt and high temperatures, that triggered a series of initial rock falls, which subsequently destabilized the surrounding rock (Guthrie et al., 2012). The flow inundated the entire valley and made its way into the mainstem Lillooet River Valley. It stopped just short of the Lillooet River campground,

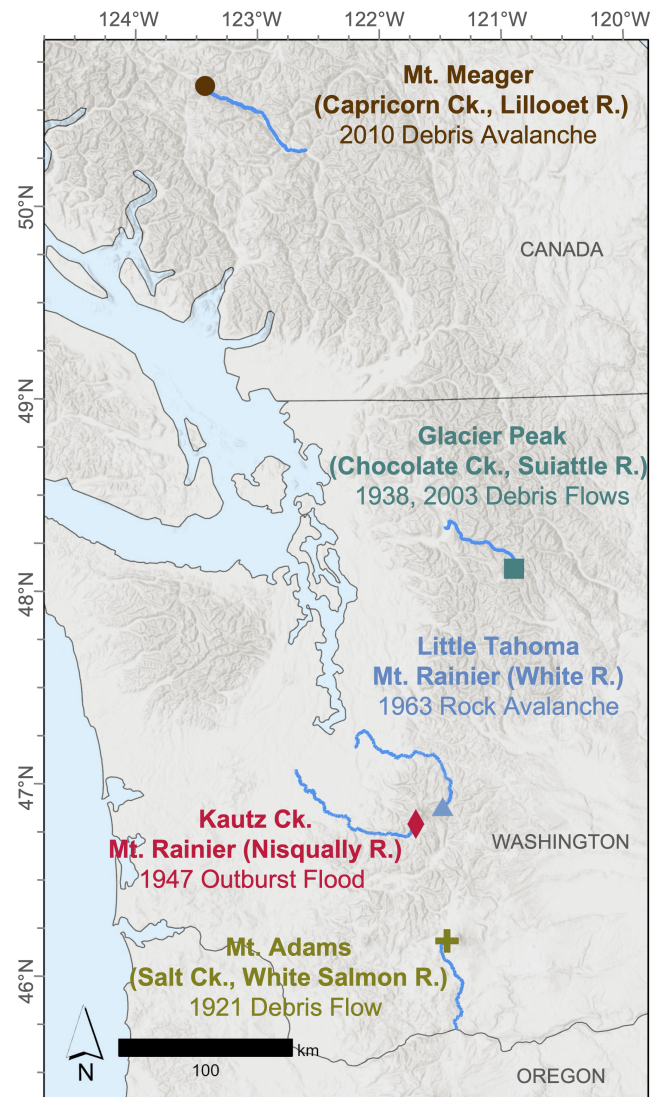


Figure 1 Overview of the 5 study deposit locations in the Pacific Northwest of North America. The associated mainstem river for each deposit is traced in blue.

roughly 2 km downstream of the Meager Creek-Lillooet River confluence, for a total runout of 12.7 km (Roberti et al., 2017). The initial failure was estimated to be $48.5 \times 10^6 \text{ m}^3$ (Guthrie et al., 2012) and to have reached an estimated average velocity of 48 m/s (Roberti et al., 2017). The flow acted in two phases: a water-rich hyperconcentrated flow that moved faster and extended further, and a water-poor debris avalanche that moved slower with discreet shear zones, forming hummocky deposits (Fig. 2f) (Roberti et al., 2017). These two end members exist on a rheological continuum that was present throughout the event. We focused on characterizing the distal and medial areas of the 2010 deposit due to physical accessibility for sampling (Fig. 2f).

2.2 Glacier Peak (Suiattle River)

Glacier Peak (3,213 m) is a stratovolcano situated in the Snoqualmie National Forest, 100 km northeast of Seattle, Washington (Figs. 1 and 2b). Several episodes of eruptions in the past 14,000 years have produced extensive volumes of lahars, pyroclastic flows, ash fall, pumice lapilli fall, and rounded cobble and boulder

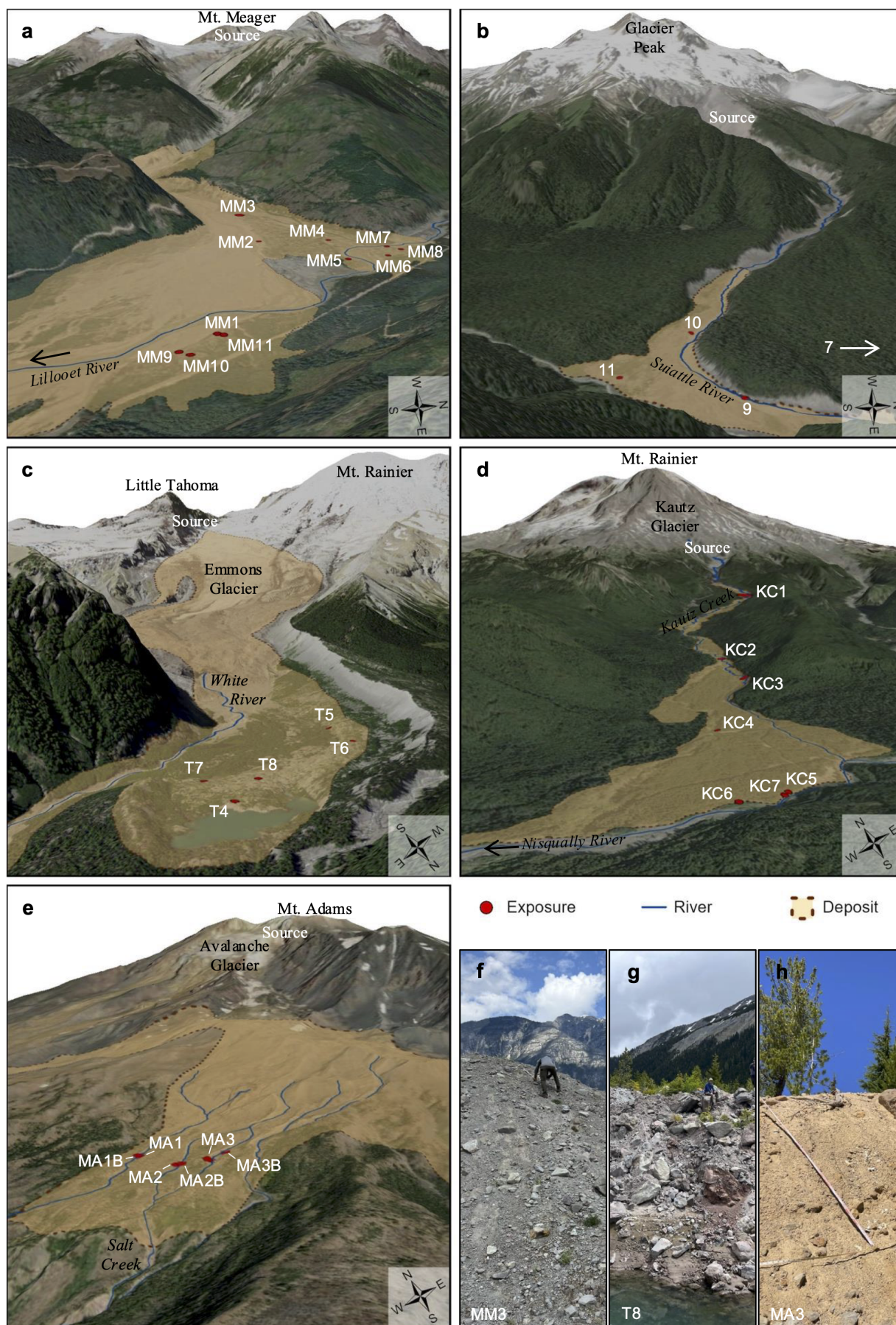


Figure 2 Overviews of the five study deposits with exposure locations (a-e). Example deposit exposures for Mt. Meager (f), Little Tahoma (g), and Mt. Adams (h).

Table 1 Key characteristics of the large volcanic mass wasting events included in this study

Mountain (source area)	Mainstem River	Year	Key citations	Source zone	Triggering event
Mt. Meager (Capricorn Ck.)	Lillooet R	2010	Guthrie et al. (2012) ; Roberti et al. (2017)	hydrothermally altered rock face	Rockfall onto glacier, channelized runout
Glacier Peak (Chocolate Ck.)	Suiattle R.	ca. 1938	Slaughter (2004)	proglacial	glacial outburst(?), entrained lahar and pyroclastic deposits
Glacier Peak (Chocolate Ck.)	White R.	1963	Crandell and Fahnestock (1965)	hydrothermally altered rock face	rockfall onto glacier, non-channelized
Mt. Rainier	Nisqually R.	1948	Crandell (1971) ; Scott et al. (1995)	proglacial	cloudburst, glacial toe collapse/outburst
Mt. Adams	White Salmon R.	1921	Vallance (1999)	hydrothermally altered rock face	rockfall onto the glacier

deposits across the landscape ([Slaughter, 2004](#)). Most of its eruptive products are dacitic in composition. In ca. 1938, a debris flow travelled 20 km down the Suiattle River. It likely originated from a glacial outburst flood at Chocolate Glacier in the headwaters of Chocolate Creek on Glacier Peak, entraining material and creating a slurry. This event was likely associated with a period of warming between 1892 and 1938 that caused significant glacial recession ([Slaughter, 2004](#)). Since the late 1930's, a total of at least nine valley-filling debris flows have occurred in the Suiattle basin with a mean recurrence interval of ca. 9.3 years (individual intervals range between 2-16 years) ([Fordham et al., 2023](#)). In 2003, another lahar filled the valley below Chocolate Creek ([Fordham et al., 2023](#)). [Pfeiffer et al. \(2022\)](#) characterized the 2003 deposit sediment (Fig. 2b). We compare these data to data collected from our other field areas.

2.3 Little Tahoma Peak at Mt. Rainier (White River)

Little Tahoma Peak is 3,395 m high and is situated on the eastern flank of Mt. Rainier, WA, towering above Emmons Glacier and the White River Valley (Figs. 1 and 2c). The peak is comprised of andesitic breccia and lava flows that have been heavily hydrothermally altered ([Crandell and Fahnestock, 1965](#)). Steep cliffs straddle Little Tahoma, a result of glacial scouring and undercutting. On December 14, 1963, part of Little Tahoma fell into the valley as a rock avalanche, covering the lower Emmons Glacier and filling the upper White River. The slide came to a stop nearly 7 km down valley near the White River Campground and the Emmons Glacier terminal moraine (Fig. 2c). The rock avalanche volume was estimated at around $10.7 \times 10^6 \text{ m}^3$ of material, with the deposit covering a total area of around 5.2 km^2 ([Crandell and Fahnestock, 1965](#)). These authors hypothesized that the rock avalanche was primed by freeze-thaw cycles and snowmelt, weakening the rock over time, and was ultimately triggered by a steam explosion at Little Tahoma Peak. This explosion weakened support and destabilized adjacent rock masses, re-

sulting in the collapse. We sampled the subvertical exposures and crumbling megaclasts in distal portions of the deposit (Fig. 2g). ([Crandell and Fahnestock, 1965](#)).

2.4 Mt. Rainier (Kautz Creek)

Kautz Glacier and Kautz Creek are situated on the southwestern flank of Mt. Rainier (4,392 m) and flow into the Nisqually River (Figs. 1 and 2d). This valley has an extensive history of lahars, outburst floods, and hyper-concentrated floods. On October 1st and 2nd of 1947, a total of 149 mm of rain doused the area, an event with a recurrence interval of 100-500 years ([Crandell, 1971](#); [Scott et al., 1995](#)). This resulted in a high magnitude runoff and surface discharge, which undercut the glacier, causing masses of ice to topple. Blocks of ice created temporary dams in the box canyon below Kautz Glacier until they collapsed and released outburst floods that bulked to debris flows in three separate pulses that surged down the valley, ultimately entraining and depositing $38 \times 10^6 \text{ m}^3$ of material in the valley ([Crandell, 1971](#); [Scott et al., 1995](#)). These debris flows had a consistency of wet concrete and carried vegetation and andesite boulders as large as 4 meters in diameter. Like many mass wasting events, this was made up of a complex series of movement styles, ranging from dilute outburst flood to debris flow with successive flow fronts that were overtaken by faster, more dilute hyper-concentrated flows ([Scott et al., 1995](#)). Due to the event dynamics, it is likely that the material was largely sourced from morainal streambed sediment. At the base of Kautz Creek valley, a fan formed due to aggradation, which inundated a valley width of 0.9 km at the highway and terminated at the Nisqually River (Fig. 2d) ([Crandell, 1971](#); [Scott et al., 1995](#)). The material from this deposit was easily identifiable in contrast to other lahar-runout deposits due to its prevalent darker, gray color, which contrasts with the lighter, more yellow colors typical of volcanic alteration present in other events in that valley.

2.5 Mt. Adams (Salt Creek)

Mt. Adams (3,743 m) is situated in southern Washington State (Figs. 1 and 2e) and is a large stratovolcano made up of heavily hydrothermally altered andesitic lava and breccia towers resulting from Quaternary volcanism (Vallance, 1999). There is a long history of lahars and debris avalanches on the southwest side of Mt. Adams, which drains into the White Salmon River. These events are a result of two key processes. Firstly, weak hydrothermally altered rock has formed steep slopes that are exposed to freeze-thaw weathering, rendering them unstable and prone to collapse (Vallance, 1999). Additionally, glacial ice scouring and plucking, followed by deposition, have formed moraines of large clasts primed to move down slope as lahars due to summer glacial ice melt and snowmelt (Vallance, 1999). There have been at least eight recorded lahars and debris avalanches in the upper reaches of the White Salmon River valley. Notably, a 1921 debris avalanche estimated to be $4 \times 10^6 \text{ m}^3$ in volume initiated at the head of Avalanche Glacier on the southwest flank of the peak, running out as much as 2 km down the Salt Creek Valley (within the White Salmon basin) and leaving a deposit ca. 1-4 m thick (Vallance, 1999). The 1921 deposit is andesitic in composition but contains breccia, scoria, and massive rocks that are hydrothermally altered with colors ranging from purple to yellow and white (Fig. 2h). These colors are indicative of andesitic rock altered to opal, alunite, kaolinite, and montmorillonite (Vallance, 1999; Finn et al., 2007).

3 Methods

Field and lab data from each of the five deposits were used to constrain the variability in sediment characteristics between volcanic mass wasting deposits in the Pacific Northwest and to calculate their abrasion potential during downstream fluvial transport. Glacier Peak field measurements were obtained from prior research (Pfeiffer et al., 2022). We conducted field campaigns at the remaining four deposits in summer 2023.

3.1 Field Measurements

At each field site, we recorded lithologic variability and GSD of the deposits, and quantified clast rock strength using a Schmidt Hammer (Schmidt Hammer Rock Strength, SHRS). We collected data from five to eleven exposures at each deposit, distributed, to the extent possible, laterally and longitudinally across the deposit, to capture potential spatial variability. Ideal exposure sites were sub-vertical scarps of the deposit, often found along river cutbanks. We chose these exposures to increase the confidence that the exposures represent the original deposits, rather than reworked material or newer sediments overlaying the target deposits.

At each exposure, we laid out a grid using stadia rods and tape measures to give 100 evenly spaced sampling locations (see Fig. 2h). Within the grid, we selected the clast underlying each nodal point to obtain a randomized, representative sample of 100 clasts for grain size,

lithology, and rock strength measurements. The clast (or a subsample of the clast) at every 5th nodal point was collected, for a total of 20 clast samples per exposure, to quantify abrasion rate and rock density in the lab. These sample clasts were selected randomly regardless of size and whether or not SHRS measurements were taken.

In some instances, it was not possible to collect ca. 100 samples as intended. For example, at the Mt. Meager deposit, we sampled only ca. 55 clasts at three exposures, MM4, 10, and 11 (Fig. 2a). These were hummocks with relatively uniform fine material. Since these hummocks contributed relatively little coarse sediment relative to the other portions of the deposit, we did not weight these measurements more heavily, despite the smaller number of measurements. Additionally, at the Kautz Creek deposit, we sampled <100 clasts at the KC1, 2, 3, 4, 5, and 7 exposures (Fig. 2d) due to exposure size and safety concerns.

For each selected clast, we measured the intermediate b-axis diameter to the nearest 5 mm using a tape measure. For embedded clasts that were too large to excavate, best judgement was used to approximate the b-axis diameter. Clasts <5 mm were classified as “fines”, and lithology was not identified.

The lithology classification scheme was specific to each field site and was determined collectively by the field team after a thorough examination of the representative lithologies present in the area prior to data collection. These classifications emphasize characteristics pertinent to abrasion potential, such as vesicularity and weathering, not petrology (Table S1). The classification system is grouped into six umbrella categories: Vesicular Altered (VA), Vesicular (V), Non-Vesicular Altered (NA), Non-Vesicular (N), Metamorphic (M), and Plutonic (P) (Table S1).

Previous work at Glacier Peak measured the strength of clasts in the deposit using a Proceq Rock Schmidt Type L (Pfeiffer et al., 2022). We followed this methodology to quantify the strength of the remaining four deposits. We determined the median SHRS of the first 50 clasts encountered with a b-axis diameter >11 cm, per exposure, though we ultimately excluded the strength data from clasts of b-axis <25 cm from our analysis, as we found that these smaller (and especially small c-axis) grains appeared to have systemically, and presumably spuriously, lower SHRS values (see Demirdag et al. (2009)). The median strength for each clast was derived from 10 measurements per clast, taken at spatially distributed planar surfaces around the clast to account for variability in strength due to fractures, local weathering, different crystals, and other sources of spatial heterogeneity (Bernard et al., 2019). The largest clasts were measured in-situ, while smaller clasts were commonly pulled from the outcrop and stabilized against a firm surface for measurements. If an exposure had fewer than 50 clasts >11 cm at the gridded sampling locations, we measured a random selection of clasts in the debris apron directly below the exposure until SHRS was recorded for 50 clasts.

3.2 Lab Work: Abrasion Tumbler and Density

Of the sample clasts brought back from the field, 34 sample clasts were used in our abrasion experiments. Chosen samples had field SHRS values and had initial clast b-axis >25 cm (prior to subsampling for transportation back from the field to the lab). Samples were selected to represent the full range of lithology types and SHRS values across all deposits, rather than an equal number from each deposit. When large enough (ca. >8 cm), we cut the field clast samples into new subsamples to obtain replicates. Due to the small size of many of the sample clasts collected in the field, only 18 of the 34 tumbled clasts have replicates.

As in O'Connor et al. (2014), we used a QT12 Lortone Rotary Rock Tumbler (4.3 L capacity) to simulate rock abrasion during bedload transport and measure the rate of mass loss of 34 clasts. We closely followed the methodology used by (Pfeiffer et al., 2022). In brief, the rock tumbler was filled with 1,890 g of host sediment, 2 L of water, and 3-6 sample clasts. The size distribution of the “host” GSD followed that of Pfeiffer et al. (2022) to simulate the gravel size and lithology mixture of the rivers downstream of our study deposits. Volcanic clasts were tumbled in a plutonic host, and plutonic samples were tumbled in a volcanic host to simulate the mixed-lithology conditions in the river and allow us to easily distinguish the sample clasts from the host. Selected clasts were cut down to <6 cm samples that would fit in the tumbler and were put through an initial rounding phase to remove the sharp angles from the clasts that tend to abrade more rapidly than rounded clasts (Domokos et al., 2014). Once rounded, the abrasion rates of the clasts were obtained during a second phase of tumbling. To calculate this, we convert tumbler experimental time to transport distance by multiplying the inner circumference of the tumbling barrel (60 cm) by the tumbler revolution rate (33 rotations per minute) and tumbler experimental time. The total tumbling distance varied depending on rock strength, ranging from 15-60 km, with 3-6 measurements of clast mass taken at intervals throughout the run. We then used the transport distance and mass loss of the clasts to determine a best-fit regression to calculate a tumbler-derived mass loss coefficient, α , in the Sternberg (1875) equation for exponential grain mass loss downstream:

$$M_x = M_0 e^{-\alpha x} \quad (1)$$

where M_0 is the original grain mass, M_x is the grain mass at a given distance downstream, x is the downstream distance, and α represents the abrasion rate, defined for a particular lithology or, in our case, individual clast.

We utilized a Mettler Toledo density determination kit, following standard operating procedures, to obtain the density for each clast (Mettler-Toledo, 2020). Rock fragments ca. 1-4 cm in diameter were cut off larger field-collected sample clasts and then were soaked in water for 48 hours prior to measuring in order to fill air bubbles before density measurements. We measured the density of all 34 clasts that were used for tumbling and an additional 120 clasts, approximately 30 per deposit, that were selected at random, distributed evenly

across all exposures at each deposit. Finally, as in Pfeiffer et al. (2022) we compare the experimentally determined clast abrasion rates and density values to measured SHRS to derive the empirical best-fit relationships for each.

3.3 Sediment Abrasion Model Analysis

To quantify the impacts of source deposit variability on downstream river channel response, we apply the downstream abrasion modeling approach of Pfeiffer et al. (2022) to our five study sites. Abrasion down a river is commonly modeled using Eq. 1, which relates mass loss to distance of downstream transport and a given abrasion rate (α). The abrasion rate coefficient, α , is often established via laboratory experiments, such as in an abrasion tumbler, as we did in this study. Though still commonly used (e.g., O'Connor et al., 2014), this simple equation neglects many other factors, such as clast fragmentation (Kodama, 1994), velocity of transport (Attal and Lavé, 2009), abrasion that occurs when a stationary rock is impacted by passing bedload (Prancevic et al., 2020) or jostled in place (Schumm and Stevens, 1973), and expedited abrasion related to initial clast angularity (Domokos et al., 2014). The standard implementation of Eq. 1 also fails to account for the fact that abrasion tumblers can underestimate the abrasion rate of a clast by overestimating the effective transport distance (Attal and Lavé, 2009).

To account for the effects of transport-dependent abrasion on the downstream loss of coarse sediment in the Suiattle River, Pfeiffer et al. (2022) took a modeling approach that we adopt here for all our study deposits. We summarize this approach below, but refer the reader to Pfeiffer et al. (2022) for further details.

The approach, developed by Pfeiffer et al. (2022), based on Attal and Lavé (2009), accounts for variations in abrasion via a baseline abrasion rate (α_{baseline}) and a transport-dependent abrasion (κ) factor using:

$$\alpha = \alpha_{\text{baseline}} \left(1 + \kappa \frac{\rho_s - \rho_w}{\rho_w} D \left(\frac{\tau^*}{\tau_c^*} - 3.3 \right) \right) \quad (2)$$

where ρ_s is the density of sediment (kg/m^3), ρ_w is the density of water (1000 kg/m^3), D is the diameter of the grain (m), τ_c^* is the critical Shields stress, τ^* is the Shields stress associated with a given grain at a given flow. Together, $\frac{\tau^*}{\tau_c^*}$ is referred to as the transport stage.

The transport stage is used to account for transport velocity dependent variation in abrasion whereas grains with a transport stage greater than roughly 3.3 abrade faster (Attal and Lavé, 2009).

The baseline abrasion (α_{baseline}) accounts for abrasion tumblers over-estimating transport distance by using a tumbler correction factor:

$$\alpha_{\text{baseline}} = \alpha_{\text{initial}} \times C_t \quad (3)$$

where C_t is the tumbler correction factor, and α_{initial} is obtained from the tumbler measurements (see below). The calculated α_{baseline} from Eq. 3 is then plugged into Eq. 2, which is used in Eq. 1 to calculate downstream mass loss.

As in Pfeiffer et al. (2022), we use our field-based SHRS measurements to estimate α_{initial} distributions for each deposit (as in section 3.2). However, the lithologic distribution of SHRS values we collected for each deposit (56-194 measurements per site) does not necessarily match the lithologic distribution represented within the entirety of each deposit, as quantified via the point counts (225-893 measurements per site). To address this, we generated a simulated dataset with 1,000 samples of SHRS for a given deposit, distributing the samples according to the lithology fractions of the whole deposit. The strengths were sampled based on an empirical distribution of the original data, ensuring the new simulated dataset reflects the distribution of the original data for each lithology and strength. Note that the distribution of vesicular and non-vesicular volcanic lithologies modeled for Glacier Peak differs from the ratio used in Pfeiffer et al. (2022). The authors used the ratio of lithologies from the boulder SHRS data, while here we use the ratio measured in the whole deposit to maintain consistency with the data from the other deposits. Thus, our resulting model for Glacier Peak, while similar, is not identical to the results presented in Pfeiffer et al. (2022).

Pfeiffer et al. (2022) implement Eqs. 1-3 in a 1-D numerical model that estimates the loss of coarse sediment from bedload to wash load, downstream due to abrasion, initial grain size, and lithologic strength and density. The model does not capture selective transport, downstream inputs of sediment, or the morphodynamic evolution of sediment pulses. Initial input parameters include the grain size, lithology, and rock strength.

The dimensionless bed shear stress associated with a given particle at a given flow, τ^* , is calculated using:

$$\tau^* = \frac{\tau}{(\rho_s - \rho_w)gD} \quad (4)$$

where g is the acceleration due to gravity (9.81 m/s^2), and τ is the dimensional shear stress at a given distance downstream.

To determine representative values of shear stress (τ) for a full channel profile, we approximate values following Pfeiffer et al. (2022). Generally, a channel profile and representative flow depth or a river-scale hydraulic model would be needed to determine τ . Instead, Pfeiffer et al. (2022) approximated the conditions of the Suiattle River by first assuming that the upstream median grain size (D_{050}) is 0.14 m (based on field measurements) and that grain size exponentially fines downstream with a diameter loss coefficient (δ) of 0.01 (based on field measurements of grain size along the length of the channel). They then calculate the median grain size at a given distance downstream (D_{x50}):

$$D_{x50} = D_{050}e^{-\delta x} \quad (5)$$

These grain diameters are then used to calculate τ for each given distance downstream:

$$\tau_x = 2\tau_c^*(\bar{\rho}_s - \rho_w)gD_{x50} \quad (6)$$

where τ_c^* is the critical dimensionless bed shear stress, assumed to be 0.045.

We used a modified version of this model to quantify sediment abrasion in the five study areas with 100 km of downstream transport. Instead of estimating a separate downstream shear stress profile for each site's longitudinal profile, we used the Suiattle boundary conditions across all the rivers, as the goal was to compare the effects of abrasion, not the channel profile, on each deposit. Here, we chose to use $k = 15$, as this value does well to match the slope of the transport-velocity-dependent abrasion data in Attal and Lavé (2009). We set $C_t = 2$ as prior work demonstrates that C_t should be >1 and <4 , with 2 as a reasonable middle ground Lewin and Brewer (2002). In the model, the α_{initial} and ρ_s of each clast are calculated using the best-fit equations from our abrasion and density measurements. Generally, rocks with a higher SHRS have a lower abrasion rate and a higher density. Importantly, if a clast's diameter is reduced to $<0.002 \text{ m}$ (smaller than gravel) due to abrasion, the model treats it as wash load and excludes it from further downstream abrasion calculations.

To summarize, we ran a sediment abrasion model for each of the five study deposits. For each deposit, the model run began with 1,000 'clasts'. Each clast had a unique grain size, SHRS value, lithologic classification, and density. The 1,000 clasts were simulated such that they collectively represented the overall GSD, SHRS distribution, and lithology distribution we observed in that deposit. All the clasts were then simulated through 100 km of downstream transport. During transport, these clasts are subject to abrasion, essentially via exponential decay (Eq. 1), with the abrasion rate of each clast modified (Eq. 2) based on clast strength, and transport stage, which is based on grain size, density, and reach shear stress (variable downstream for a given clast; Eq. 6). As the clasts abrade, the grain size of each clast reduces accordingly, and all other parameters stay the same. If a clast's diameter is reduced to $<0.002 \text{ m}$, it is treated as wash load and no longer tracked, leaving only the clasts $>0.002 \text{ m}$. This also allows us to track how the volumetrically weighted lithology mixtures and mean SHRS of the coarse load in each deposit change over the course of the 100 km of downstream transport.

4 Results

4.1 Lithology

The five study deposits all originate from stratovolcanoes in the Pacific Northwest with intermediate composition. However, the degree of vesicularity and alteration of clast types varies substantially between deposits (Fig. 3, Table S1). The Kautz Creek deposit is almost entirely composed of non-vesicular rock (dark purple; Fig. 3d), in contrast to the other deposits, which all have $>50\%$ vesicular material. The Glacier Peak and Kautz Creek deposits exhibit no alteration (shades of orange; Figs. 3b, d), whereas the Mt. Adams deposit is ca. 60% hydrothermally altered (Fig. 3e). It is worth noting that for Little Tahoma, the vesicular material contains the full spectrum of alteration (like Mt. Adams), but this was not accounted for in the lithology classification scheme set for that deposit (Table S1). Despite all hav-

ing source zones on volcanic edifices, three of the mass wasting deposits contain plutonic material and notably, Mt. Meager contains ca. 30% plutonic rock, as well as the only occurrence of metamorphic rock (Fig. 3a). Within each deposit, lithologic trends are generally consistent across exposures, although some variability is present (Table S2; Pinke, 2024).

4.2 Rock Strength

As with the lithology categories, the five study deposits contain a broad range of rock strength distributions. The SHRS distributions are significantly different between deposits ($p = 1.19 \times 10^{-11}$), with mean SHRS ranging from 43.5 to 58.4 (Fig. 4, Table S3). Rock strengths are generally consistent across the sampling exposures within each of the deposits (Fig. S1), but are highly variable for each given lithologic classification. Rock strength measurements are lacking for the hummock exposures in the Mt. Meager deposit and the distal exposures in the Kautz Creek deposit, due to the absence of larger clasts (Figs. 4b, e). Across all the study deposits (Fig. 4f), clasts that are vesicular and hydrothermally altered are weaker, whereas rock that is non-vesicular and not altered is stronger. The lithologic fractions for each deposit (Fig. 3) are mapped to the strength distributions for each lithology (Fig. 4) as a modeled lithology-weighted distribution of rock strengths for each deposit (left-hand distributions, Figs. 4a-e). The Kautz Creek deposit at Mt. Rainier, which primarily consists of non-vesicular, non-altered material, has the highest mean strength of 58.4 (Fig. 4d), and is followed by clasts at Glacier Peak, which also contain no hydrothermally altered clasts (Fig. 4b). Mt. Adams, which has a high percentage of altered and vesicular material (70.7% combined) is the weakest of the deposits with a mean strength of 43.5 (Fig. 4e).

4.3 Grain Size

The grain size of hillslope material entering a river channel may vary due to rock strength, weathering and erosion history, and emplacement style (Sklar et al., 2017; Roda-Boluda et al., 2018). The five mass wasting deposits exhibit a range of GSDs (Fig. 5). The D_{16} , D_{50} , and D_{95} are within the same order of magnitude between deposits, but within each deposit, the GSD spans two orders of magnitude, varying from fine sand to meter-scale boulders (Fig. 5, Table S4). The Glacier Peak and Kautz Creek deposits have the highest modeled SHRS (Figs. 4b, d, Table S3) and the largest D_{50} (9 cm) and second largest D_{50} (7 cm, Figs. 5c, e, Table S4), whereas the weakest deposit, Mt. Adams, also has the smallest D_{50} (1.5 cm; Fig. 5f). The fraction of fine material (<5 mm) varies drastically between deposits, from <20% to >50%, with Kautz Creek exhibiting the highest percent of fines (Figs. 3 and 5, Table S4). For most of the deposits, we find that the individual exposures have similar GSDs to one another. This lends confidence that the GSD we characterize is representative of the whole deposit, despite the modest number of exposures. Furthermore, at Mt. Adams, we sampled exposures on both

sides of three small valleys, and in each case, the GSD of the adjacent exposures on the two valley walls match up well (Fig. 5f). Kautz Creek is the exception, with substantial differences in grain size between exposures. The Kautz Creek data exhibit a trend with larger grain sizes at exposures proximal to the source and within the main body and finer grain sizes in the distal portions of the deposit (Fig. 5e). The Mt. Meager and Little Tahoma deposits depict GSD differentiation between types of exposures. In both instances, the GSD of hummocks are finer, whereas the GSD of scarps of the main matrix of the deposit are coarser (Figs. 5b, d).

4.4 Abrasion and Density

The subset of clasts selected for abrasion experiments spanned three orders of magnitude in calculated abrasion rates based on tumbling (Fig. 6a). Density of the clasts from this subset spanned a range of >1000 kg/m³ (Fig. 6b). Pfeiffer et al. (2022) related SHRS to clast abrasion rate and density. We added 34 clasts from all of our deposits to their dataset and found that although the SHRS-to-abrasion-rate and SHRS-to-density relationships are still significant, the strength of the relationships is more modest than Pfeiffer et al. (2022) found using rocks only from Glacier Peak. Intuitively, weaker and less dense rocks abrade faster, whereas stronger, denser rocks abrade slower (Fig. 6). Differences in the mean density between deposits are marginally non-significant ($p = 0.059$). Like the lithology-weighted SHRS distribution (Fig. 4), the median weighted density for Mt. Meager, Little Tahoma, and Mt. Adams is similar, while Kautz Creek is higher (see Pinke, 2024).

4.5 Downstream Model

We used the lithologies for each deposit (Fig. 3), their corresponding SHRS-proxy abrasion rates (Figs. 4 and 6), and their GSDs (Fig. 5) to model downstream abrasion in terms of coarse sediment volumetric loss (Fig. 7). The Glacier Peak and Kautz Creek deposits exhibited the least amount of abrasion over a simulated 100 km, with 56% and ca. 52% loss, respectively. These two deposits have the strongest clasts, with material exhibiting little to no hydrothermal alteration and a large fraction of non-vesicular rock. In contrast, the Mt. Meager, Little Tahoma, and Mt. Adams deposits experience substantially more abrasion downstream. The Mt. Adams deposit, which is the most hydrothermally altered and begins with the lowest mean SHRS, loses ca. 86% of its coarse volume to wash load over a simulated 100 km. The loss can also be considered through the lens of the 'halving distance' of the pulse material: the transport distance over which half of the coarse sediment is lost to abrasion. Among our sites, this distance varies from 13 km (Mt. Adams) to 92 km (Kautz Creek; Fig. 7f). In each deposit, most of the volumetric material loss occurs in the weaker lithologies and in the first 20 km downstream, whereas the stronger lithologies persist in greater relative quantities downstream. Across all deposits, the mean SHRS increases downstream, with deposits of lower starting SHRS generally associated with

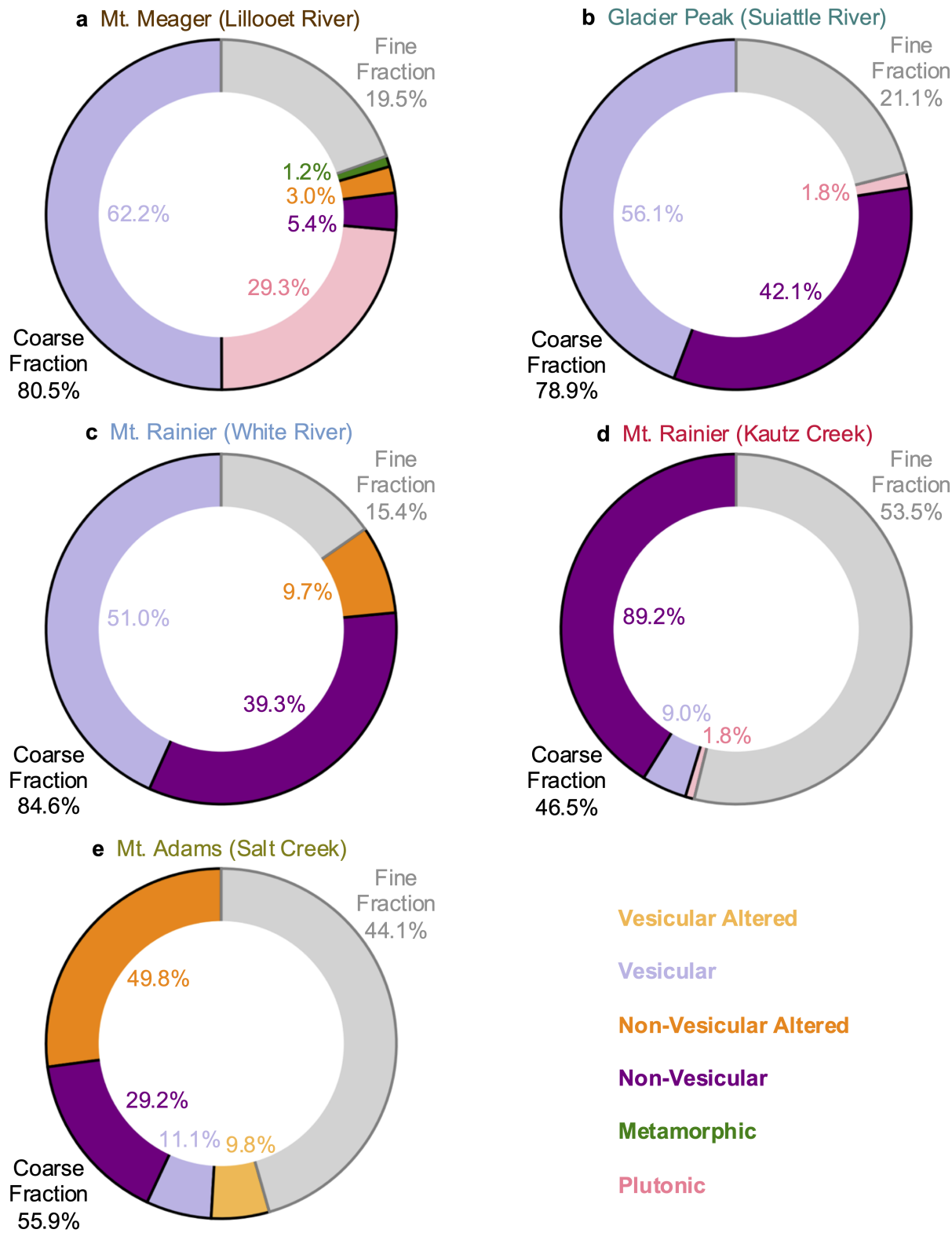


Figure 3 Lithologic percents for the coarse (>5 mm) material fraction of each deposit. Clasts <5 mm are denoted as fines in gray.

a greater change in strength over the 100 km of modeled transport (Fig. 7f). On average, the remaining material is 9.4% stronger (in SHRS units) than the initial, but ranges from 4.2% (Kautz Creek) to 14.2% (Mt. Meager). Mt. Meager represents a notable case, with a bimodal starting distribution in rock strength (Fig. 4a) leading to the greatest downstream change in mean SHRS among the sites (Fig. 7f). This results from the rapid loss of vesicular volcanics, leaving a downstream coarse sedi-

ment pulse dominated by the durable surviving plutonics. This result highlights both the large magnitude of the abrasion effect in some volcanic deposits and the importance of the lithologic mixture in setting the pattern of downstream loss.

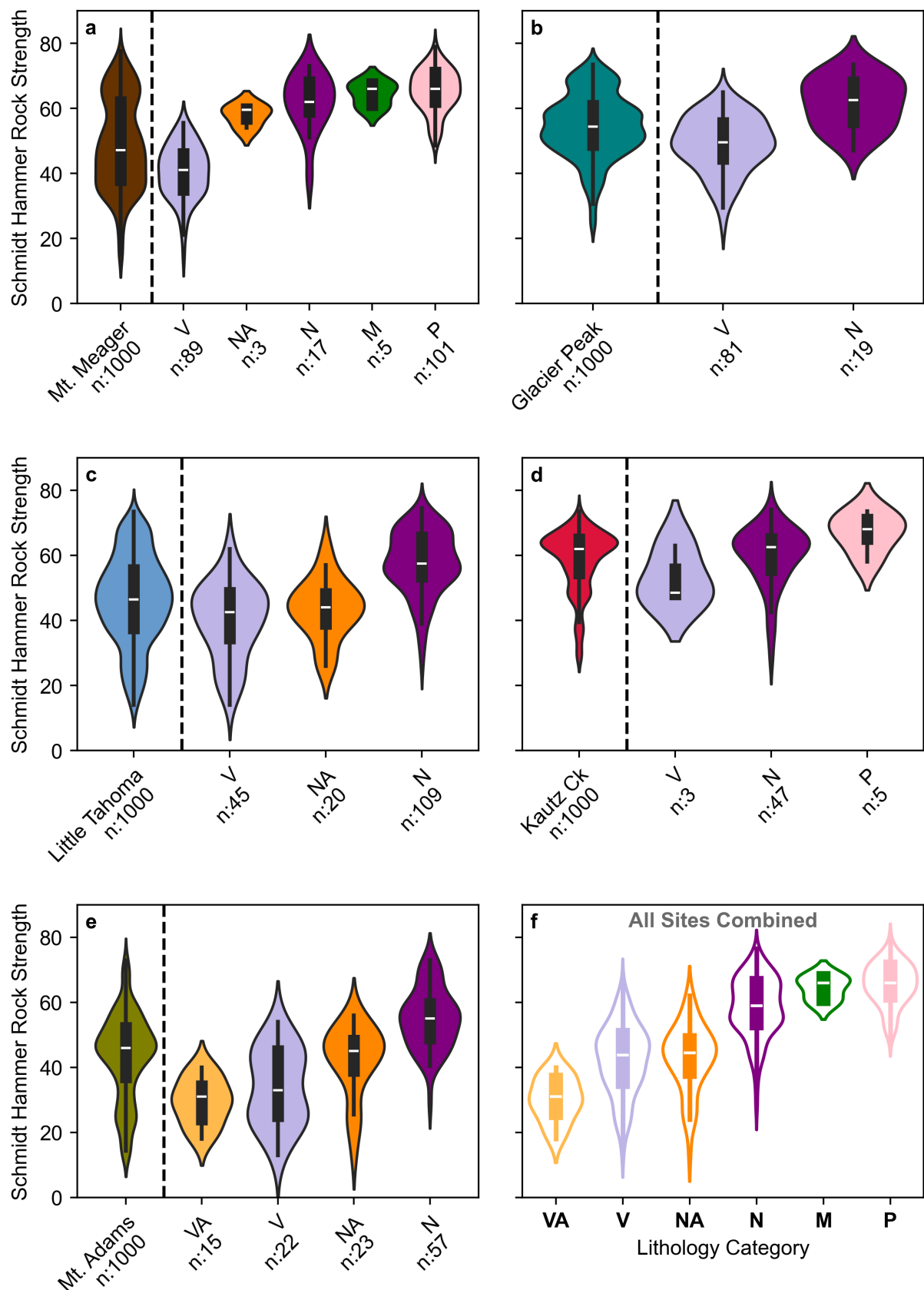


Figure 4 Schmidt Hammer Rock Strength (SHRS) of each deposit by lithology type (a-e), and for each lithologic category, combined across sites (f). Deposit distribution is weighted to lithologic fraction (Figure 3) at the far left of each panel. Colors as in Figure 1 (sites) and Figure 3 (lithology categories). VA: vesicular altered, V: vesicular, NA: non-vesicular altered, N: non-vesicular, M: metamorphic, P: plutonic.

5 Discussion

Substantial variation in mass wasting sediment strength and grain size between our study deposits leads to dra-

matic differences in their durability downstream. Kautz Creek and Mt. Adams represent two ends of the spectrum (Fig. 7). With sediments dominated by coarse,

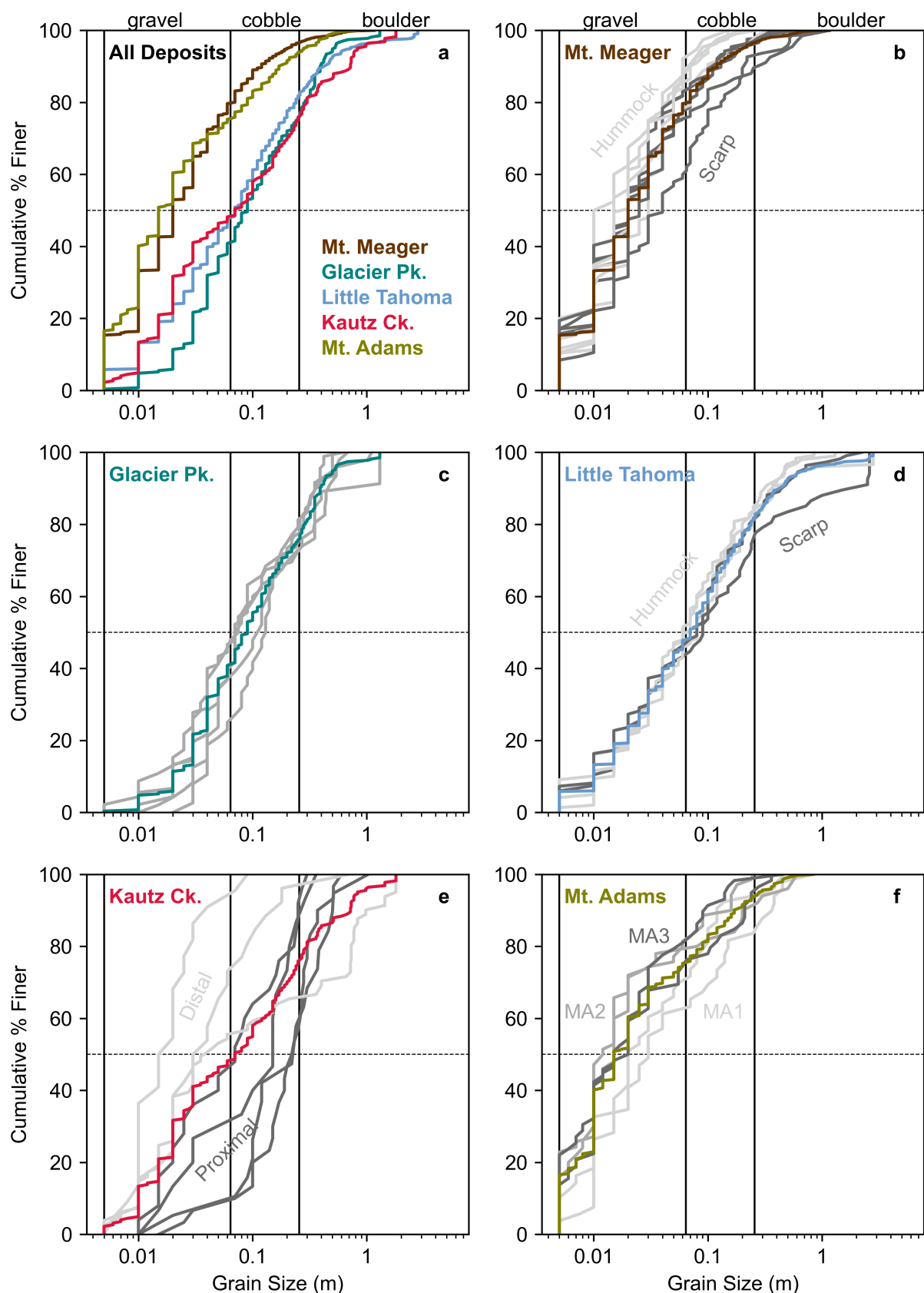


Figure 5 Coarse (>5 mm) fraction grain size distribution of all deposits (a) and deposits with each exposure (b-f), excluding clasts <5 mm. Individual exposures in gray.

strong, unaltered, non-vesicular rock, nearly half of the Kautz Creek material persists through 100 km of simulated abrasion. In contrast, the Mt. Adams sediments, predominantly made up of finer-grained, weaker, al-

tered, vesicular rock, attenuate dramatically in 10s of km downstream of the source zone. In all cases, the surviving material downstream strengthens compared to the initial pulse.

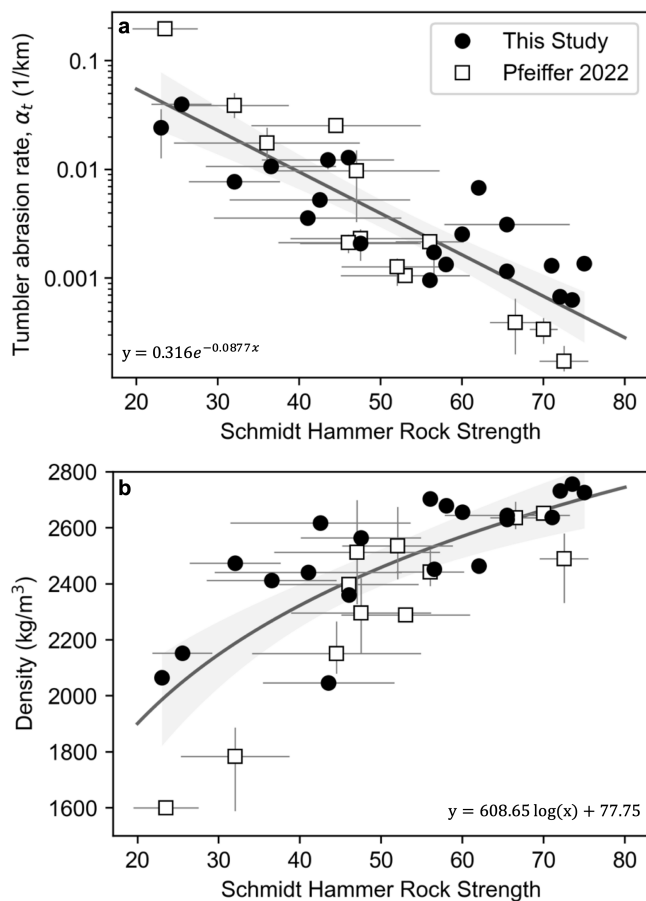


Figure 6 (a) Schmidt Hammer Rock Strength (SHRS) vs. abrasion rate. (b) SHRS vs. Density. Trendline is for this study, and Pfeiffer et al. (2022) combined datasets.

Below, we examine the initial GSDs of our study deposits and the influence of rock strength on these grain sizes. Then we explore the potential for similarities between deposits on a single volcanic edifice and the sensitivity of our downstream abrasion model to the sediment pulse GSD. We then expand our abrasion model framework to a global compilation of rock strength to consider how sediment pulse abrasion at our sites may compare to other rock types. Finally, we consider implications of downstream sediment abrasion on cascading hazards.

5.1 Grain Size

A source-to-sink model for landscape evolution must capture the grain size dynamics from hillslope to channel (Sklar, 2024). The median grain size of our five study deposits falls within the reported ranges for hillslope sediment supply from previous studies (Neely and DiBiase, 2020; Roda-Boluda et al., 2018). However, the widths of our measured GSDs are notable. The volcanic sediment contributed to these study channels ranges in size from silt to boulders, with D_{84}/D_{50} ratios from 3.8 to 8.0 (Mt. Meager and Mt. Adams, respectively) (Table S5). These are wider than most of the grain size distributions reported in the literature coming from (non-volcanic) landslides and talus piles, with coarse-fraction D_{84}/D_{50} ratios ranging from 1.3 (chert, Verdian et al., 2021) to a maximum of 5.1 (granitic, channel head San

Gabriel Mtns, Neely and DiBiase, 2020). While sampling schemes may explain some portion of this discrepancy (studies handle the fine fraction differently, and fine sediment may be systematically missing from the surface of talus piles due to kinetic sieving), we suggest that our study deposits may have systematically wider GSDs than previous studies due to differences in the process of emplacement and/or clast creation.

In steep bedrock landscapes, the grain size distribution of talus piles is well explained by the fracture spacing in the source bedrock (Neely and DiBiase, 2020; Verdian et al., 2021). In contrast, on stratovolcanoes, some portion of the pre-failure sediment at most of these sites would have been tephra deposits, with the original GSD set by the eruption dynamics and distance from the vent. In addition, all of the deposits we sampled, except Little Tahoma, represent material that transported some portion of the runout distance as a debris flow, introducing the potential for clast abrasion and therefore evolution of the GSD, during the runout (Caballero et al., 2012; Arabnia and Sklar, 2016). Finally, hydrothermal alteration present at volcanoes may weaken the rock, making it more prone to break down during emplacement and to generating clays after. This is exemplified by the Mt. Adams deposit which has the largest fraction of hydrothermally altered material (Fig. 3) and the widest D_{84}/D_{50} ratio (Table S5). Therefore, the distinct lithologic and transport histories of volcanic mass wasting deposits may result in substantially different initial grain size distributions supplied to the fluvial system.

These wide GSDs of our studied mass wasting deposits have implications for the subsequent fluvial transport of that sediment and the complexity of modeling sediment pulse motion downstream. Morphodynamic model results suggest that sediment pulses with wide GSDs increase bed mobility as compared to narrower pulse GSD's (Ahammad et al., 2021). This suggests that efforts to model the downstream transport of large mass wasting events must adequately capture not just the pulse sediment D_{50} , but the full GSD. Most gravel-bedded rivers have D_{84}/D_{50} ratios of ca. 2 (Rickemann and Recking, 2011). As a result of this common observation, when modeling the downstream transport of sediment pulses, researchers often assign the pulses a uniform or narrow GSD with D_{84}/D_{50} ratios <2 (Venditti et al., 2010; Arnaud et al., 2017; Gaeuman et al., 2017). However, in our study sites, the sediment supply does not start out as uniform. Rather, the mass wasting sediments enter the river systems as very wide GSDs with substantial heterogeneity in sediment strength and density. We can infer that the fine material is quickly carried away as part of the wash load. Of the remaining bed material, the low-strength (and as a corollary low-density; Fig. 6) material is more subject to efficient downstream transport, and therefore abrasion. These processes raise the D_{50} and thereby narrow the deposit-proximal in-channel GSD, and strengthen the bed material relative to the deposit. Sites like those in this study, with wide sediment pulse GSDs, are likely to have riverbed GSD that differs dramatically from the pulse GSD. Consequently, we suggest that accurately capturing this initial variability in the size (and strength) distri-

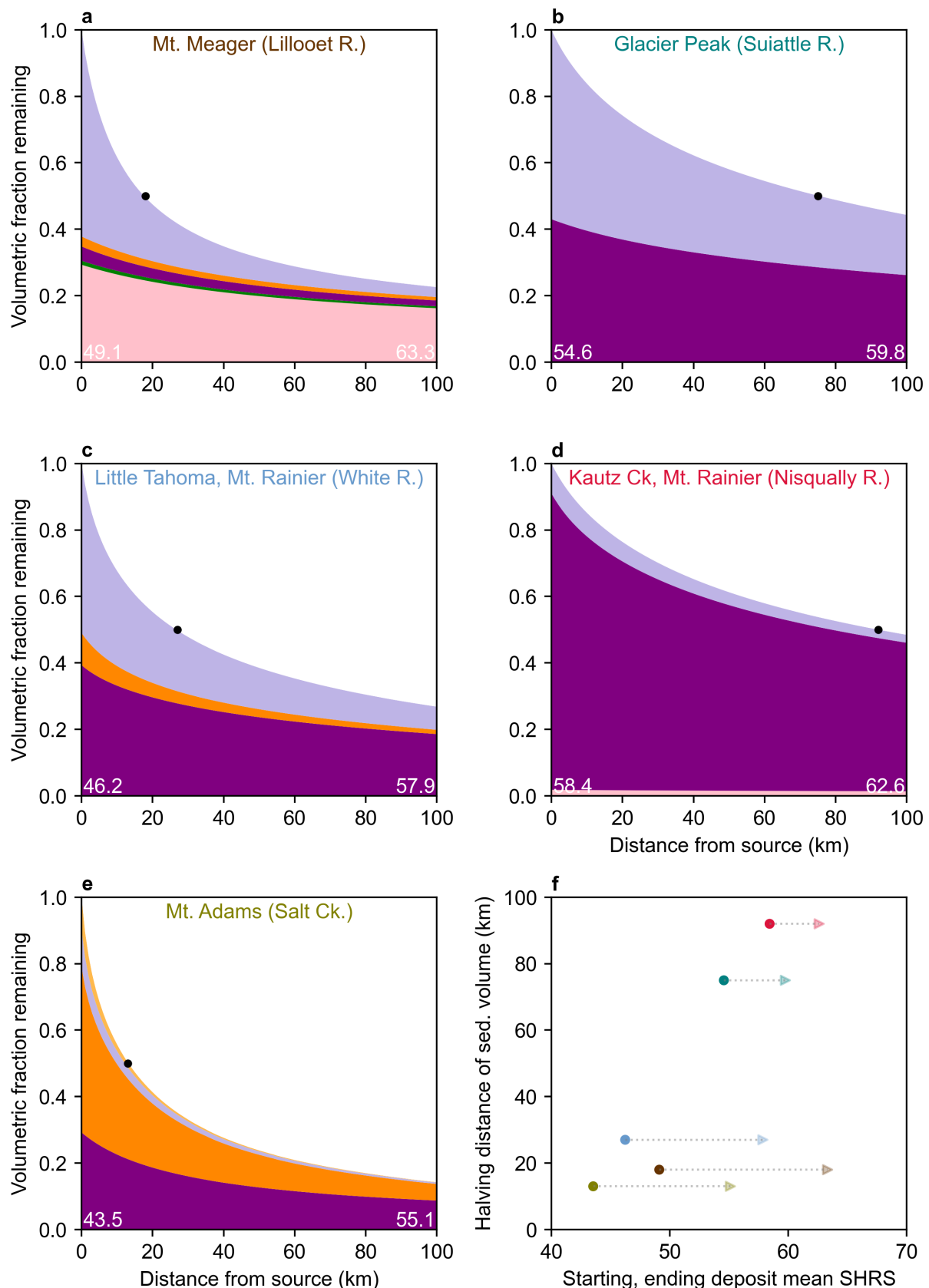


Figure 7 Modeled sediment abrasion of the five study deposits over 100 km of simulated downstream transport (a-e). The products of abrasion (wash load) are shown in white. Start and end mean SHRS of sediment are denoted in white on the left and right corners. Colors as in Figures 3 and 4. Black dots denote the distance at which half of the original coarse sediment volume is lost to abrasion. Panel (f) shows this halving distance for each of the sites plotted against the starting (circles) and 100 km model end point (triangle) mean SHRS of the coarse material. Symbols colored by deposit, as in a-e.

bution of the sediment pulse is essential for predicting the downstream evolution of the river channel following such large-scale disturbances.

5.2 Rock Strength vs. Initial Size

Prior research demonstrates that rock mass strength directly correlates to the size of clasts sourced from hill-

slopes into rivers (Roda-Boluda et al., 2018), such that stronger source rocks generally result in larger grain sizes. However, within a single lithology, GSDs vary substantially. Additionally, the degree of weathering and fracturing in the rock mass plays an important role (Roda-Boluda et al., 2018). Our data, from a very different setting, are consistent with these findings. The deposits with the strongest clasts, Glacier Peak and Kautz Creek, have the coarsest D_{50} values as compared to the GSDs of Mt. Meager and Mt. Adams, which are both weaker. Notably, the two strongest deposits hold in common a proglacial sediment source, with events triggered by icefall or outburst flood, while the other three deposits, which were sourced from the failure of a hydrothermally altered rock face. However, the deposit-scale patterns do not hold true at the grain-scale: individual sampled clast size shows no relationship to measured rock strength (Fig. S2). Thus, while individual clast strength does not relate to grain size, the previous finding of stronger hillslope bedrock yielding larger clasts is generally supported by our data from volcanic mass wasting deposits.

Little Tahoma does not tidily fit the relationship between rock strength and grain size observed by Roda-Boluda et al. (2018) and present in our other sites. The clasts we measured at Little Tahoma are weak but have a coarse D_{50} (Figs. 4c and 5d). In the field, we observed that boulders were often crumbling in place. While we recorded the size of the intact clasts, in many cases, we could pull scraps of rock apart with our bare hands. It was not clear why these rocks did not disintegrate during transport. Perhaps the clasts, despite being weak and hydrothermally altered, were able to remain as intact blocks because of unconfined runout on the Emmons Glacier. Alternatively, perhaps exposure to sub-aerial weathering processes for the 60+ years since deposit emplacement has resulted in their crumbly nature. Preliminary research using rocks from this study supports this, suggesting that exposure to just 40 wet-dry and freeze-thaw cycles can weaken these rocks (Coffin, 2024). This suggests that some volcanic lithologies may be prone to rapid weathering post-deposition, yielding low measured strengths mere decades after the mass wasting event, thus obscuring the original strength-to-size relationship.

5.3 One Volcano, Two Deposits

Our data suggest that both mass wasting failure mechanisms and source may have substantial impacts on the size distribution and strength of the deposits. It may be tempting to attribute the observed differences between the deposits to the bulk composition of the igneous rocks at the four different volcanic edifices in the Pacific Northwest, each with a unique combination of eruptive styles and rock geochemistry. However, the Little Tahoma and Kautz Creek deposits, both on Mt. Rainier, differ as well. Their source zones lie on opposite sides of Mt. Rainier, the northeast and southwest corners, respectively (Fig. 1). This provides an opportunity to parse the influence of the volcano writ large vs. the specific volcanic units the deposits are derived from

and the emplacement style, on the resulting deposit characteristics. The Little Tahoma and Kautz Creek deposits are sourced from different regions of Mt. Rainier with distinctly different hydrothermal alteration histories and event styles. Whereas the Little Tahoma material was sourced from a heavily hydrothermally altered satellite peak on the flank of Mt. Rainier, the Kautz Creek material was sourced from unaltered andesite from rock falls, or rock plucked off Mt. Rainier during glaciation that was resting as morainal streambed sediment (Scott et al., 1995). The Kautz Creek material is considerably stronger than the Little Tahoma material (median clast SHRS 62 vs. 46; Fig. 4, Table S3), and the outburst flood nature of Kautz Creek likely lends itself to the higher fines fraction as compared to Little Tahoma (Fig. 3). The stark differences between these two deposits, sourced from the same volcanic edifice, caution against assuming that mass wasting deposits of the same volume, from the same mountain, will have uniform potential impacts downstream.

5.4 Uncertainties in Downstream Abrasion Simulation

While the downstream abrasion simulations (Fig. 7) presented here provide useful estimates of the coarse sediment loss due to abrasion, they are subject to some uncertainty. Given that the Mt. Adams deposit material has the finest D_{50} and experiences the most volumetric material loss, and similarly, the Glacier Peak deposit has the coarsest D_{50} and experiences the second-lowest amount of volumetric loss, it is worth querying whether the abrasion model results are largely a function of the GSD of each deposit. This is possible, as our model incorporates transport-rate enhanced abrasion (Attal and Lavé, 2009) that increases the per-km abrasion rate of small, and therefore highly mobile, grains. To test this, we ran the model for two additional scenarios using the Mt. Adams (finest) and Glacier Peak (coarsest) GSDs along with the associated lithologies and strengths for each of the five deposits (Figs. S3 and S4). Kautz Creek had the largest difference in volumetric loss between the big vs. small GSD model scenarios (ca. 16%), while Mt. Adams had the smallest difference (ca. 10%). In both scenarios, the strength trends hold firm, with the Glacier Peak and Kautz Creek exhibiting less volumetric loss than the other three deposits, and Mt. Adams exhibiting the most loss. Especially of note, when the Glacier Peak deposit is modelled with the smaller GSD, it retains more of its initial material (ca. 29%) than the Mt. Adams deposit does with the larger GSD (ca. 24%) (Figs. S3 and S4). These results suggest that while the abrasion model outcomes are heavily influenced by the GSD of each deposit, with finer gravel sizes abrading more rapidly, lithology and clast strength serve as the primary controls.

5.5 Extending Our Findings to General Rock Abrasion Potential

Our findings suggest that coarse sediment loss to abrasion can be highly efficient in the first tens of km downstream from these Pacific Northwest volcanic mass

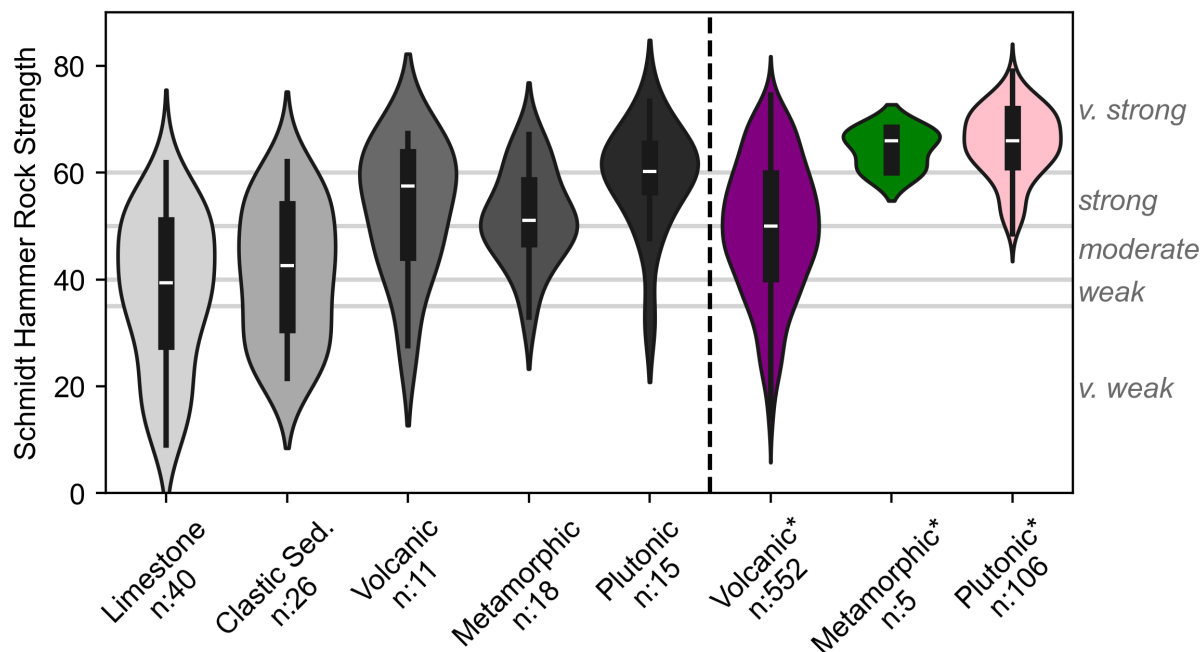


Figure 8 Compilation of Schmidt Hammer Rock Strength (SHRS) of different rock types globally, modified from Goudie (2006). Each data point (n) is the mean of many measurements of that lithology from a single study. Data from this study are denoted with an asterisk (*) and color. Strength categories listed in grey italics on the right are from Goudie (2006); v.: very.

wasting deposits (Fig. 7). While the source material is heterogeneous and, often, of only ‘moderate’ strength (O’Connor et al., 2014), that sediment quickly evolves into material that is consistently ‘strong’ to ‘very strong’ (Figs. 7 and 8). This work builds on that of O’Connor et al. (2014), who modeled region-scale bed material abrasion in Western Oregon, considering the implications of the rapid loss of the weakest lithologies (Coast Range sedimentary rocks). Our abrasion rates are, on average, in the same range as those reported by O’Connor and colleagues. The key advance of our approach is the incorporation of source sediment heterogeneity, which allows us to resolve the impact of survivorship bias: that within a given lithologic category, the coarse sediments that persist downstream in the fluvial system will be the strongest ones.

To assess the broader relevance of our findings to source-to-sink sediment transport in other geologic settings, we use a global compilation of SHRS values (Goudie, 2006) to compare how other lithologies and rock strength distributions may fare in downstream transport. The Goudie (2006) rock strength values compare well to the data from this study (Fig. 8), with our measured volcanic, metamorphic, and plutonic lithology types all falling within the ranges of those respective lithologies in the global compilation (Goudie, 2006).

We can use the Goudie (2006) compilation to better understand how the downstream abrasion of our volcanic sediment pulses may compare with other sediment pulse types globally (Fig. 9). In our study, the Mt. Adams deposit comprises the weakest material, with ca. 14% of the original coarse rock volume persisting after 100 km of simulated downstream transport, with the downstream survivors consisting of the strongest initial rock types (Fig. 7). When applying the

downstream abrasion model towards the global rock strengths (for more detail see Pinke, 2024), the weaker rock such as limestone and (clastic) sedimentary rock (mean SHRS = 42.2) exhibit 86% and 82% volumetric loss, respectively, over 100 km (Fig. 9), while plutonic rock (mean SHRS = 58.1) loses only 50% (Fig. 9). It is important to note that these data are from a global compilation, and thus the range of rock strengths presented for each lithology may not be representative of a single outcrop or formation. We suggest that uniform rock strengths for a given lithology (Fig. 9f) are likely the exception rather than the standard. For example, sedimentary rock strength varies between beds, rocks exposed to hydrothermal alteration are highly spatially heterogeneous, faulted rock units have varying degrees of damage, and weathering-prone rocks may enter the channel in varied states of alteration. The implications of our findings, highlighting how weak and heterogeneous rocks control the durability of mass wasting sediment pulses downstream, likely extend to many geologic settings beyond the Pacific Northwest stratovolcanoes studied here. In particular, sedimentary rocks, which are on average substantially weaker than the volcanic rocks represented by our field sites, may be particularly prone to dramatic rates of downstream abrasion (e.g., O’Connor et al., 2014).

5.6 Implications for Cascading Hazards and Flood Mitigation

Cascading hazards have been a recent area of focus in the Earth surface processes community (Yanites et al., 2025), with the aim of better understanding and ultimately forecasting the timing and magnitude of downstream impacts of large headwater geomorphic distur-

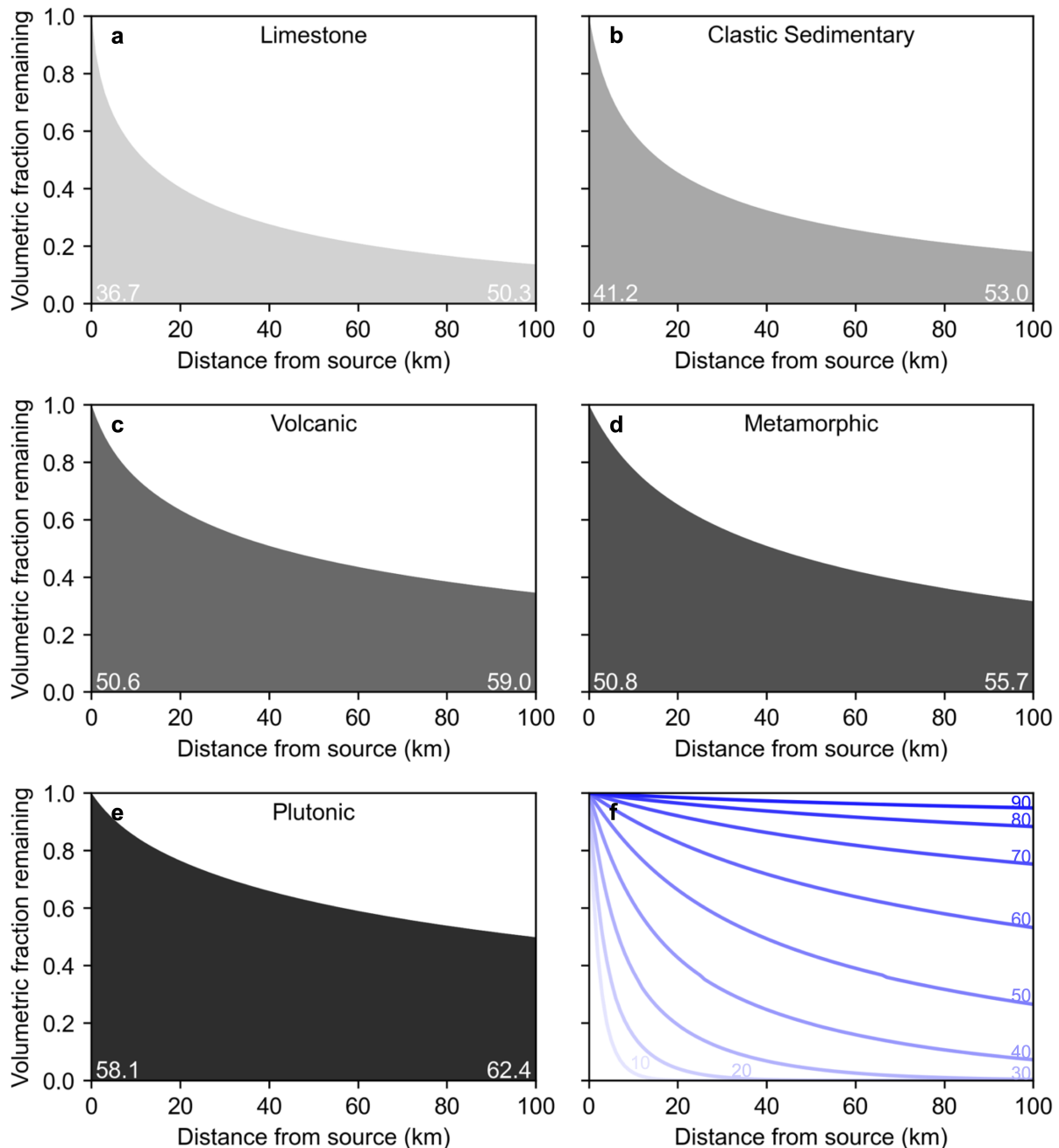


Figure 9 Modeled abrasion of pulses with global rock compilation (a-e, see Fig. 8) and with pulses of varying uniform strength, with SHRS values for each curve shown in blue (f), using the Little Tahoma GSD. The products of abrasion (wash load) are shown in white. Start and end mean Schmidt Hammer Rock Strength (SHRS) of sediment, denoted in white on the left and right corners.

bances. Climate change has the potential to increase the frequency of triggering events such as those in this study. Thawing rock permafrost was implicated in the Mt. Meager event, exceptionally wet weather preceded the Kautz Creek and Mt. Adams events, while a glacial outburst or icefall event has been suggested as the trigger at Glacier Peak. As these events increase in frequency in the coming decades, we argue that accurate predictions of hazard cascades through the fluvial system will require characterizing grain size and abrasion potential of mass wasting contributions to channel networks. Grain size acts as an important filter on the fluvial component of the hazard cascade. For example, the 2014 Oso landslide disaster, while locally devastating and deadly, had minimal effects downstream as the

failed hillslope was dominated by glacial outwash sands and glaciolacustrine silt (Wartman et al., 2016), fine material that was transported downstream as wash load. On the other hand, channel bed aggradation patterns in the gravel-bedded Nooksack River (Anderson and Konrad, 2019), hint at a decades-long downstream propagating bed wave triggered by meteorological events in the early 20th century. Our findings, building on those of Pfeiffer et al. (2022), demonstrate that lithology can act as a filter on cascading hazards, by reducing the potential for cascading downstream coarse sediment impacts in landscapes where the supplied sediment is readily abraded by fluvial processes. So, while much of the recent progress on hazard cascades has (and will continue to) come from technological advances, such as

high temporal and spatial resolution satellite imagery and advances in modeling, our findings demonstrate the necessity of field data to capture not only the GSD but also the abrasion potential of the material.

Accurately predicting the propagation of bed sediment waves has financial and human implications. The management response to the 2010 Mt. Meager event may be an example of the risk of ignoring the effect of abrasion. Our data suggest that abrasion may be substantially attenuating (Fig. 7a) the downstream propagation of the sediment pulse currently passing down the Lillooet River toward the community of Pemberton, British Columbia, Canada. Flood hazards increased substantially in the years immediately following the event, with 0.4–2 m of aggradation through the agricultural and urban mid- and lower river (beginning ca. 40 km downstream from the deposit; (Northwest Hydraulic Consultants, 2018). In a report written 8 years after the Mt. Meager event, consultants working on behalf of the Pemberton Valley Dyking District reported that, while sand and the finest gravel passed quickly downstream to the (populated) lower river, the leading edge of the coarse “sediment wave” was still propagating down the upper river, and should be expected to cause further future increases in flood hazard. They recommend the removal of >200,000 m³ of sediment annually from the channel for the next ca. 20 years to mitigate potential increased flood hazards (Northwest Hydraulic Consultants, 2018). They base this interpretation, in part, on the lithologies observed in the bed material on bars down the length of the channel. They use the relative concentration of a distinctive porphyritic rhyodacite (ca. 60% of the deposit, correlative to vesicular, V, in our classification scheme, which we measure as 62% of the deposit, Fig. 3a) as a tracer, indicative of the 2010 mass wasting material. They report that the concentration of the porphyritic rhyodacite decreases over the first ca. 25 km downstream of the deposit, and is nearly absent by 60 km downstream. Based on our modeling (Fig. 7a), we expect this ‘tracer’ should disappear from the gravel bars downstream at approximately the rate they observed via abrasion, irrespective of the rate of transport of a bed wave. This is to say that vesicular volcanics are not conservative tracers. This oversight has implications for hazard mitigation. If we assume that the products of abrasion transport as wash load, the flood managers are potentially overestimating the magnitude of costly upper river sediment removal required to mitigate future increases to flood hazards, as up to ca. 50% of the looming wave of bed material will likely abrade away by the time it arrives to the populated lower river. Neglecting the process of bed material abrasion is common practice in geomorphic modeling, but in places like Mt. Meager we argue that it plays a significant role in the partitioning between fine and coarse sediment, and therefore in the hazard cascade.

6 Conclusion

Volcanic terrain contributes disproportionately to catastrophic mass wasting events in the Pacific Northwest. These sediment pulses have a large impact on

the proximal reaches of river systems, but the coarse sediment pulse signal may attenuate downstream due to sediment abrasion. The effectiveness of abrasion in attenuating the signal depends on the grain size distribution of the original deposit, as well as its lithologic makeup. We have constrained the sediment characteristics of five Pacific Northwest mass wasting deposits and found significant heterogeneity between them. Sediment pulses similar to Kautz Creek, predominantly made up of stronger, unaltered, non-vesicular rock, have a strong potential to persist over 100 km of downstream transport. In contrast, pulses like Mt. Adams, predominantly made up of weaker, altered, vesicular rock, are likely to attenuate dramatically in 10s of km. In all cases, the surviving material downstream strengthens compared to the initial pulse.

The implications of this work can be extrapolated more broadly, for other lithologies and strengths, when considering the overall sediment budget of a river and downstream flood hazard management. We show that sedimentary rocks, in particular, may be prone to abrasion to an extent beyond that of our volcanic sites. The importance of downstream abrasion may be easy to miss when bed sediment samples are taken only from the riverbed in reaches far from the hillslope source, where there is survivorship bias of the strongest rocks. Future efforts could use the data presented here as inputs to a morphodynamic model of downstream sediment transport and the effects of abrasion, predicting temporal changes in bed elevation and surface grain size.

Acknowledgements

We are grateful to the many field assistants who made this work possible: Charlie Vryhof, Devin Johnson, Ashanie Long-Reid, H. Hoffer, Olivia Pedersen, Fate-meh Shacheri, Josh Sayre, Colby Rand, and Edward Mina. Thanks to the members of the Fluvial Lab for feedback on the manuscript, and to Ben Paulson for steadfast logistical support. We thank two anonymous reviewers and the associate editor for thoughtful and constructive feedback that improved the quality of this contribution, as well as Ciro Cerrone for copyediting.

Land Recognition

This study focused on the volcanoes and rivers of the ancestral homelands of the Sauk-Suiattle, Puyallup, and Nisqually Tribes, the Klickitat people of the Yakima Nation, and the Lílwat Nation. We express our gratitude for their continued stewardship of these lands and waterways.

Data and Codes Availability

The data and code that reproduce the analysis in this study are archived on Zenodo at: <https://doi.org/10.5281/zenodo.17316266>.

Funding Statement

This work was funded by the National Science Foundation (GLD Award #2138504 to A. Pfeiffer) and the Geological Society of America Graduate Student Research Grant to B. Pinke.

Conflict of Interest Disclosure

The authors have no competing interests.

Permission to Reproduce Material from Copyrighted Sources

The authors declare that no material from copyrighted sources was reproduced in this manuscript.

Declaration on Artificial Intelligence Use

A Large Language Model was used to assist with Python scripting to troubleshoot bugs in the code and to improve the presentation of several figures, as well as to check the accessibility of the non-technical summary. Beyond that, the authors declare that no artificial intelligence tool was used to generate any part of the contents in this manuscript.

References

- Ahammad, M., Czuba, J. A., Pfeiffer, A. M., Murphy, B. P., and Belmont, P. Simulated dynamics of mixed versus uniform grain size sediment pulses in a gravel-bedded river. *Journal of Geophysical Research: Earth Surface*, 126(10):e2021JF006194, 2021. doi: [10.1029/2021JF006194](https://doi.org/10.1029/2021JF006194).
- An, C., Cui, Y., Fu, X., and Parker, G. Gravel-bed river evolution in earthquake-prone regions subject to cycled hydrographs and repeated sediment pulses. *Earth Surface Processes and Landforms*, 42(14):2426–2438, 2017. doi: [10.1002/esp.4195](https://doi.org/10.1002/esp.4195).
- Anderson, S. W. and Konrad, C. P. Downstream-propagating channel responses to decadal-scale climate variability in a glaciated river basin. *Journal of Geophysical Research: Earth Surface*, 124(4):902–919, 2019. doi: [10.1029/2018JF004734](https://doi.org/10.1029/2018JF004734).
- Arabnia, O. and Sklar, L. S. Experimental study of particle size reduction in geophysical granular flows. *International Journal of Erosion Control Engineering*, 9(3):122–129, 2016. doi: [10.13101/ijece.9.122](https://doi.org/10.13101/ijece.9.122).
- Arnaud, F., Piégay, H., Béal, D., Collery, P., Vaudor, L., and Rollet, A.-J. Monitoring gravel augmentation in a large regulated river and implications for process-based restoration. *Earth Surface Processes and Landforms*, 42(13):2147–2166, 2017. doi: [10.1002/esp.4161](https://doi.org/10.1002/esp.4161).
- Attal, M. and Lavé, J. Pebble abrasion during fluvial transport: Experimental results and implications for the evolution of the sediment load along rivers. *Journal of Geophysical Research: Earth Surface*, 114(F4), 2009. doi: [10.1029/2009JF001328](https://doi.org/10.1029/2009JF001328).
- Bernard, T., Sinclair, H. D., Gailleton, B., Mudd, S. M., and Ford, M. Lithological control on the post-orogenic topography and erosion history of the Pyrenees. *Earth and Planetary Science Letters*, 518:53–66, 2019. doi: [10.1016/j.epsl.2019.04.034](https://doi.org/10.1016/j.epsl.2019.04.034).
- Caballero, L., Sarocchi, D., Borselli, L., and Cárdenas, A. I. Particle interaction inside debris flows: Evidence through experimental data and quantitative clast shape analysis. *Journal of Volcanology and Geothermal Research*, 231–232:12–23, 2012. doi: [10.1016/j.jvolgeores.2012.04.007](https://doi.org/10.1016/j.jvolgeores.2012.04.007).
- Coffin, F. Effects of Experimental Freeze-Thaw and Wet-Dry Weathering Cycles on Fluvial Abrasion Rates, 2024. https://cedar.www.edu/wwu_honors/857.
- Crandell, D. R. Postglacial lahars from Mount Rainier Volcano, Washington, 1971, <https://doi.org/10.3133/pp677>.
- Crandell, D. R. and Fahnestock, R. K. Rockfalls and avalanches from Little Tahoma Peak on Mount Rainier, Washington, 1965, <https://doi.org/10.3133/b1221A>.
- Cui, Y., Parker, G., Lisle, T. E., Gott, J., Hansler-Ball, M. E., Pizzuto, J. E., Allmendinger, N. E., and Reed, J. M. Sediment pulses in mountain rivers: 1. Experiments. *Water Resources Research*, 39(9), 2003. doi: [10.1029/2002WR001803](https://doi.org/10.1029/2002WR001803).
- Davies, T., Phillips, C., and Warburton, J. Processes, transport, deposition, and landforms: Flow. In *Treatise on Geomorphology*, edited by Shroder, J. F., volume 7, pages 158–170. Elsevier, 2013. doi: [10.1016/B978-0-12-374739-6.00160-3](https://doi.org/10.1016/B978-0-12-374739-6.00160-3), <https://doi.org/10.1016/B978-0-12-374739-6.00160-3>.
- del Potro, R. and Hürlimann, M. The decrease in the shear strength of volcanic materials with argillic hydrothermal alteration, insights from the summit region of Teide stratovolcano, Tenerife. *Engineering Geology*, 104(1–2):135–143, 2009. doi: [10.1016/j.enggeo.2008.09.005](https://doi.org/10.1016/j.enggeo.2008.09.005).
- Demirdag, S., Yavuz, H., and Altindag, R. The effect of sample size on Schmidt rebound hardness value of rocks. *International Journal of Rock Mechanics and Mining Sciences*, 46(4):725–730, 2009. doi: [10.1016/j.ijrmms.2008.09.004](https://doi.org/10.1016/j.ijrmms.2008.09.004).
- Domokos, G., Jerolmack, D. J., Sipos, A. Á., and Török, Á. How river rocks round: Resolving the shape-size paradox. *PLOS ONE*, 9(2): e88657, 2014. doi: [10.1371/journal.pone.0088657](https://doi.org/10.1371/journal.pone.0088657).
- Erdmer, P. and Cui, Y. Geological map of British Columbia (Geoscience Map 2009-1), 2009, https://cmscontent.nrs.gov.bc.ca/geoscience/PublicationCatalogue/GeoscienceMap/BCGS_GM2009-01.pdf.
- Finn, C. A., Sisson, T. W., and Deszcz-Pan, M. Aerogeophysical measurements of collapse-prone hydrothermally altered zones at Mount Rainier volcano. *Nature*, 409:600–603, 2001. doi: [10.1038/35054533](https://doi.org/10.1038/35054533).
- Finn, C. A., Deszcz-Pan, M., Anderson, E. D., and John, D. A. Three-dimensional geophysical mapping of rock alteration and water content at Mount Adams, Washington: Implications for lahar hazards. *Journal of Geophysical Research: Solid Earth*, 112(B5), 2007. doi: [10.1029/2006JB004783](https://doi.org/10.1029/2006JB004783).
- Fordham, E. M., Pfeiffer, A. M., Bunn, A. G., and Novak, S. J. Recurrent debris flows and their downstream fate: Geomorphic drivers of an anomalous sediment load, Suittale River, Washington State, USA. *Earth Surface Processes and Landforms*, 48(6):1307–1320, 2023. doi: [10.1002/esp.5551](https://doi.org/10.1002/esp.5551).
- Gaeuman, D., Stewart, R., Schmandt, B., and Pryor, C. Geomorphic response to gravel augmentation and high-flow dam release in the Trinity River, California. *Earth Surface Processes and Landforms*, 42(14):2523–2540, 2017. doi: [10.1002/esp.4191](https://doi.org/10.1002/esp.4191).
- Geertsema, M., Clague, J. J., Schwab, J. W., and Evans, S. G. An overview of recent large catastrophic landslides in northern British Columbia, Canada. *Engineering Geology*, 83(1–3): 120–143, 2006. doi: [10.1016/j.enggeo.2005.06.026](https://doi.org/10.1016/j.enggeo.2005.06.026).
- Goudie, A. S. The Schmidt Hammer in geomorphological research. *Progress in Physical Geography: Earth and Environment*, 30(6): 703–718, 2006. doi: [10.1177/0309133306071954](https://doi.org/10.1177/0309133306071954).
- Guthrie, R. H., Friele, P., Allstadt, K., Roberts, N., Evans, S. G., Delaney, K. B., Roche, D., Clague, J. J., and Jakob, M. The 6 August 2010 Mount Meager rock slide-debris flow, Coast Moun-

- tains, British Columbia: Characteristics, dynamics, and implications for hazard and risk assessment. *Natural Hazards and Earth System Sciences*, 12(5):1277–1294, 2012. doi: [10.5194/nhess-12-1277-2012](https://doi.org/10.5194/nhess-12-1277-2012).
- Guzzetti, F., Peruccacci, S., Rossi, M., and Stark, C. P. Rainfall thresholds for the initiation of landslides in central and southern Europe. *Meteorology and Atmospheric Physics*, 98(3–4): 239–267, 2007. doi: [10.1007/s00703-007-0262-7](https://doi.org/10.1007/s00703-007-0262-7).
- Handwerger, A. L., Fielding, E. J., Sangha, S. S., and Bekaert, D. P. S. Landslide sensitivity and response to precipitation changes in wet and dry climates. *Geophysical Research Letters*, 49(14): e2022GL099499, 2022. doi: [10.1029/2022GL099499](https://doi.org/10.1029/2022GL099499).
- Humphries, R., Venditti, J. G., Sklar, L. S., and Wooster, J. K. Experimental evidence for the effect of hydrographs on sediment pulse dynamics in gravel-bedded rivers. *Water Resources Research*, 48(1), 2012. doi: [10.1029/2011WR010419](https://doi.org/10.1029/2011WR010419).
- Hutter, K., Svendsen, B., and Rickenmann, D. Debris flow modeling: A review. *Continuum Mechanics and Thermodynamics*, 6(1): 1–35, 1994. doi: [10.1007/BF01175749](https://doi.org/10.1007/BF01175749).
- Iverson, R. M. The physics of debris flows. *Reviews of Geophysics*, 35(3):245–296, 1997. doi: [10.1029/97RG00426](https://doi.org/10.1029/97RG00426).
- John, D. A., Sisson, T. W., Breit, G. N., Rye, R. O., and Vallance, J. W. Characteristics, extent and origin of hydrothermal alteration at Mount Rainier Volcano, Cascades Arc, USA: Implications for debris-flow hazards and mineral deposits. *Journal of Volcanology and Geothermal Research*, 175:289–314, 2008. doi: [10.1016/j.jvolgeores.2008.04.004](https://doi.org/10.1016/j.jvolgeores.2008.04.004).
- Kodama, Y. Experimental study of abrasion and its role in producing downstream fining in gravel-bed rivers. *Journal of Sedimentary Research*, 64(1a):76–85, 1994. doi: [10.2110/jsr.64.76](https://doi.org/10.2110/jsr.64.76).
- Kuenen, H. Experimental abrasion of pebbles: 2. Rolling by current. *The Journal of Geology*, 64:336–368, 1956. doi: [10.1086/6236370](https://doi.org/10.1086/6236370).
- LaHusen, S. R., Duvall, A. R., Booth, A. M., Grant, A., Mishkin, B. A., Montgomery, D. R., Struble, W., Roering, J. J., and Wartman, J. Rainfall triggers more deep-seated landslides than Cascadia earthquakes in the Oregon Coast Range, USA. *Science Advances*, 6(41):eaba6790, 2020. doi: [10.1126/sciadv.aba6790](https://doi.org/10.1126/sciadv.aba6790).
- Lewin, J. and Brewer, P. A. Laboratory simulation of clast abrasion. *Earth Surface Processes and Landforms*, 27(2):145–164, 2002. doi: [10.1002/esp.306](https://doi.org/10.1002/esp.306).
- Major, J. J., Pierson, T. C., and Scott, K. M. Debris flows at Mount St. Helens, Washington, USA. In *Debris-flow hazards and related phenomena*, edited by Jakob, M. and Hungr, O., pages 685–731. Springer, 2005. doi: [10.1007/3-540-27129-5_27](https://doi.org/10.1007/3-540-27129-5_27), https://doi.org/10.1007/3-540-27129-5_27.
- Mettler-Toledo. *Density kit reference manual for advanced and standard analytical balances*, 2020.
- Neely, A. B. and DiBiase, R. A. Drainage area, bedrock fracture spacing, and weathering controls on landscape-scale patterns in surface sediment grain size. *Journal of Geophysical Research: Earth Surface*, 125(10):e2020JF005560, 2020. doi: [10.1029/2020JF005560](https://doi.org/10.1029/2020JF005560).
- Northwest Hydraulic Consultants. Lillooet River floodplain mapping, 2018.
- O'Connor, J. E., Mangano, J. F., Anderson, S. W., Wallick, J. R., Jones, K. L., and Keith, M. K. Geologic and physiographic controls on bed-material yield, transport, and channel morphology for alluvial and bedrock rivers, western Oregon. *Geological Society of America Bulletin*, 126(3–4):377–397, 2014. doi: [10.1130/B30831.1](https://doi.org/10.1130/B30831.1).
- Pfeiffer, A. M., Morey, S., Karlsson, H. M., Fordham, E. M., and Montgomery, D. R. Survival of the strong and dense: Field evidence for rapid, transport-dependent bed material abrasion of heterogeneous source lithology. *Journal of Geophysical Research: Earth Surface*, 127(3):e2021JF006455, 2022. doi: [10.1029/2021JF006455](https://doi.org/10.1029/2021JF006455).
- Pinke, B. Vanishing Volcanics: Mass Wasting Deposit Heterogeneity, with Implications for Rapid Downstream Abrasion, 2024, <https://cedar.wvu.edu/wwuet/1340/>.
- Prancevic, J. P., Sklar, L. S., and Dietrich, W. E. Modeling how rivers digest boulders: Long residence times and coarse riverbed surfaces lead to preferential wear of boulders through in-place abrasion. In *AGU Fall Meeting Abstracts*, volume 2020, pages EP007–05, 2020.
- Read, P. B. Mount Meager complex, Garibaldi Belt, southwestern British Columbia. *Geoscience Canada*, 17(3):167–170, 1990.
- Rickenmann, D. and Recking, A. Evaluation of flow resistance in gravel-bed rivers through a large field data set. *Water Resources Research*, 47(7), 2011. doi: [10.1029/2010WR009793](https://doi.org/10.1029/2010WR009793).
- Rickenmann, D., Laigle, D., McArdell, B. W., and Hübl, J. Comparison of 2D debris-flow simulation models with field events. *Computational Geosciences*, 10(2):241–264, 2006. doi: [10.1007/s10596-005-9021-3](https://doi.org/10.1007/s10596-005-9021-3).
- Roberti, G., Friele, P., van Wyk de Vries, B., Ward, B., Clague, J. J., Perotti, L., and Giardino, M. Rheological evolution of the Mount Meager 2010 debris avalanche, southwestern British Columbia. *Geosphere*, 13(2):369–390, 2017. doi: [10.1130/GES01389.1](https://doi.org/10.1130/GES01389.1).
- Roda-Boluda, D. C., D'Arcy, M., McDonald, J., and Whittaker, A. C. Lithological controls on hillslope sediment supply: Insights from landslide activity and grain size distributions. *Earth Surface Processes and Landforms*, 43(4):956–977, 2018. doi: [10.1002/esp.4281](https://doi.org/10.1002/esp.4281).
- Schaefer, L. N., Kereszturi, G., Kennedy, B. M., and Villeneuve, M. Characterizing lithological, weathering, and hydrothermal alteration influences on volcanic rock properties via spectroscopy and laboratory testing: a case study of Mount Ruapehu volcano, New Zealand. *Bulletin of Volcanology*, 85:43, 2023. doi: [10.1007/s00445-023-01657-w](https://doi.org/10.1007/s00445-023-01657-w).
- Schraml, K., Thomschitz, B., McArdell, B. W., Graf, C., and Kaitna, R. Modeling debris-flow runout patterns on two alpine fans with different dynamic simulation models. *Natural Hazards and Earth System Sciences*, 15(7):1483–1492, 2015. doi: [10.5194/nhess-15-1483-2015](https://doi.org/10.5194/nhess-15-1483-2015).
- Schumm, S. A. and Stevens, M. A. Abrasion in place: A mechanism for rounding and size reduction of coarse sediments in rivers. *Geology*, 1(1):37–40, 1973. doi: [10.1130/0091-7613\(1973\)1<37:AIPAMF>2.0.CO;2](https://doi.org/10.1130/0091-7613(1973)1<37:AIPAMF>2.0.CO;2).
- Schuster, J. E. Geologic map of Washington state (Geologic Map GM-53), 2005, <https://ngmdb.usgs.gov/Info/dmt/docs/schuster07b.pdf>.
- Scott, K. M., Vallance, J. W., and Pringle, P. T. Sedimentology, behavior, and hazards of debris flows at Mount Rainier, Washington, 1995, <https://doi.org/10.3133/pp1547>.
- Sklar, L. S. Grain size in landscapes. *Annual Review of Earth and Planetary Sciences*, 52:663–692, 2024. doi: [10.1146/annurev-earth-052623-075856](https://doi.org/10.1146/annurev-earth-052623-075856).
- Sklar, L. S., Fadde, J., Venditti, J. G., Nelson, P., Wydzga, M. A., Cui, Y., and Dietrich, W. E. Translation and dispersion of sediment pulses in flume experiments simulating gravel augmentation below dams. *Water Resources Research*, 45(7), 2009. doi: [10.1029/2008WR007346](https://doi.org/10.1029/2008WR007346).
- Sklar, L. S., Riebe, C. S., Marshall, J. A., Genetti, J., Leclere, S., Lukens, C. L., and Mercers, V. The problem of predicting the size distribution of sediment supplied by hillslopes to rivers. *Geomorphology*, 277:31–49, 2017. doi: [10.1016/j.geomorphology](https://doi.org/10.1016/j.geomorphology).

[morph.2016.05.005](#).

- Slater, L. J., Singer, M. B., and Kirchner, J. W. Hydrologic versus geomorphic drivers of trends in flood hazard. *Geophysical Research Letters*, 42(2):370–376, 2015. doi: [10.1002/2014GL062482](#).
- Slaughter, S. L. The 1938 Chocolate Glacier debris flow, Glacier Peak volcano, North Cascades, Washington, 2004.
- Sohn, Y. K., Rhee, C. W., and Kim, B. C. Debris flow and hyperconcentrated flood-flow deposits in an alluvial fan, northwestern part of the Cretaceous Yongdong Basin, Central Korea. *The Journal of Geology*, 107(1):111–132, 1999. doi: [10.1086/314334](#).
- Sterling, S. and Slaymaker, O. Lithologic control of debris torrent occurrence. *Geomorphology*, 86(3–4):307–319, 2007. doi: [10.1016/j.geomorph.2006.09.002](#).
- Sternberg, H. Untersuchungen über Längen- und Querprofil geschiebeführender Flüsse. *Zeitschrift für Bauwesen*, 25: 483–506, 1875.
- Stoffel, M. and Huggel, C. Effects of climate change on mass movements in mountain environments. *Progress in Physical Geography: Earth and Environment*, 36(4):421–439, 2012. doi: [10.1177/0309133312441010](#).
- Sutherland, D. G., Hansler Ball, M., Hilton, S. J., and Lisle, T. E. Evolution of a landslide-induced sediment wave in the Navarro River, California. *Geological Society of America Bulletin*, 114(8):1036–1048, 2002. doi: [10.1130/0016-7606\(2002\)114<1036:EOALIS>2.0.CO;2](#).
- Vallance, J. W. Postglacial lahars and potential hazards in the White Salmon River system on the southwest flank of Mount Adams, Washington, 1999, <https://doi.org/10.3133/b2161>.
- Venditti, J. G., Dietrich, W. E., Nelson, P. A., Wydzga, M. A., Fadde, J., and Sklar, L. Mobilization of coarse surface layers in gravel-bedded rivers by finer gravel bed load. *Water Resources Research*, 46(7), 2010. doi: [10.1029/2009WR008329](#).
- Verdian, J. P., Sklar, L. S., Riebe, C. S., and Moore, J. R. Sediment size on talus slopes correlates with fracture spacing on bedrock cliffs: Implications for predicting initial sediment size distributions on hillslopes. *Earth Surface Dynamics*, 9(4):1073–1090, 2021. doi: [10.5194/esurf-9-1073-2021](#).
- Viparelli, E., Solari, L., and Hill, K. M. Downstream lightening and upward heavying: Experiments with sediments differing in density. *Sedimentology*, 62(5):1384–1407, 2015. doi: [10.1111/sed.12187](#).
- Wartman, J., Montgomery, D. R., Anderson, S. A., Keaton, J. R., Benoît, J., de la Chapelle, J., and Gilbert, R. The 22 March 2014 Oso landslide, Washington, USA. *Geomorphology*, 253:275–288, 2016. doi: [10.1016/j.geomorph.2015.10.022](#).
- Watters, R. J., Zimbelman, D. R., Bowman, S. D., and Crowley, J. K. Rock Mass Strength Assessment and Significance to Edifice Stability, Mount Rainier and Mount Hood, Cascade Range Volcanoes. *Pure and Applied Geophysics*, 157:957–976, 2000. doi: [10.1007/s000240050012](#).
- Yanites, B. J., Clark, M. K., Roering, J. J., West, A. J., Zekkos, D., Baldwin, J. W., Cerovski-Darriau, C., Gallen, S. F., Horton, D. E., Kirby, E., Leshchinsky, B. A., Mason, H. B., Moon, S., Barnhart, K. R., Booth, A., Czuba, J. A., McCoy, S., McGuire, L., Pfeiffer, A., and Pierce, J. Cascading land surface hazards as a nexus in the Earth system. *Science*, 388(6754), 2025. doi: [10.1126/science.adp9559](#).

The article *Vanishing volcanics: Implications of mass wasting deposit sediment heterogeneity on downstream abrasion and the fluvial hazard cascade* © 2026 by Pinke et al. is licensed under a [Creative Commons Attribution 4.0 International \(CC BY 4.0\) License](#).