ABSTRACT

Volcanic eruptions fundamentally alter landscapes, paving over channels, decimating biota, and emplacing fresh, unweathered material. The fluvial incision of blocky lava flows is a geomorphic puzzle. First, high surface permeability and lack of sediment should preclude geomorphically effective surface runoff and dissection. Furthermore, past work has demonstrated the importance of extreme floods in driving incision via column toppling and plucking in columnar basalt, but it is unclear how incision occurs in systems where surface blocks are readily mobile. We examine rapid fluvial incision of the Collier lava flow, an andesitic Holocene lava flow in the High Cascades of Oregon. Since lava flow emplacement ~1600 yr ago, White Branch Creek has incised bedrock gorges up to 8 m deep into the coherent core of the lava flow and deposited >0.2 km$^3$ of sediment on the lava flow surface. Field observation points to a bimodal discharge regime in the channel, with evidence for both annual snowmelt runoff and outburst floods from Collier glacier, as well as historical evidence of vigorous glacial meltwater. To determine the range of discharge events capable of incision in White Branch Creek, we used a mechanistic model of fluvial abrasion. We show that the observed incision implies that moderate flows are capable of both initiating channel formation and sustaining incision. Our results have implications for the evolution of volcanic systems worldwide, where glaciation and/or mass wasting may accelerate fluvial processes by providing large amounts of sediment to otherwise porous, sediment-starved landscapes.

INTRODUCTION

What Sets the Pace of Fluvial Incision in Volcanic Landscapes?

Effusive volcanic eruptions reset fluvial landscapes by paving over channel networks, killing and damaging vegetation, and burying soils (Jefferson et al., 2014). Though volcanic landscape disturbance has been well documented in modern explosive eruptions (Major et al., 2000; Gran and Montgomery, 2005; Swanson et al., 2013), the processes driving the post-eruption fluvial dissection of lava flows are relatively unconstrained (Jefferson et al., 2010; Deligne et al., 2013). Blocky lava flows, with high surface permeability (Freeze and Cherry, 1979) and low nutrient content (Deligne et al., 2013), which inhibit both surface water and vegetation growth (Jefferson et al., 2014), present a particular conundrum in volcanic landscape evolution. Previous workers have proposed that permeability reduction via in situ weathering (e.g., Lohse and Dietrich, 2005) is an essential precursor to drainage development in these systems (Jefferson et al., 2010).

However, there are many prominent canyons incised into lava flows where soil development is limited or absent (Lamb et al., 2008a, 2014; Baynes et al., 2015). Investigation of these channels has demonstrated that they formed as a result of catastrophic floods in excess of the magnitude of meteorologic events (e.g., O’Connor and Beebee, 2009). These floods are often caused by dam breaches, both from natural impoundments (moraines, landslides, glacial ice; Costa and Schuster, 1988) and man-made structures (e.g., Lamb and Fonstad, 2010). Past work has documented channels that are cut into columnar basalt flows or other jointed rocks, where prominent sets of vertical joints enable rapid incision via plucking (Whipple et al., 2000a) and block toppling (Lamb and Dietrich, 2009). For these systems, smaller discharges cannot mobilize the plucked blocks or lack the tools to perform abrasion (e.g., Baynes et al., 2015) and are hence unable to incise the bed.

Columnar jointing is not a ubiquitous feature of lava flows. Blocky lava flows, associated with stratovolcanoes and volcanic fields worldwide, are composed of a rubbly surface layer and a coherent, massive core (Kilburn, 2000). Though the blocks of the surface layers can be up to several meters in diameter (Deardorff and Cashman, 2012), the lack of consolidation means that unlike in columnar flows, toppling is not required to initiate incision. Nevertheless, where blocky lava flows and alpine glaciation (or large landslides) co-occur, it is likely that lava flows are subject to large, sediment-laden discharge events in the form of dam breach floods (O’Connor and Costa, 2008). These events have the potential to provide the tools and geomorphic power necessary to carve fluvial channels, perhaps through bedrock abrasion, which is typically much slower than erosion via plucking (Whipple et al., 2000a). Determining the rate and style of incision processes due to these large-scale events is essential to understanding the evolution of temperate volcanic landscapes as well as unraveling feedbacks between surface change and hydrologic response.

Approach and Scope

Here, we focus on the fluvial incision of Collier lava flow, a basaltic andesite flow in the High Cascades of Oregon where White Branch Creek has incised up to 8 m into the flow and deposited >0.2 km$^3$ of alluvial sediment (Fig. 1; Deligne et al., 2013). Using a combination of field evidence, high-resolution light detection and ranging (LiDAR) topography, and numerical modeling, we ask: Can extreme events (sensu O’Connor and Costa, 2008) achieve rapid inci-
To predict incision rates along White Branch Creek, we used a one-dimensional (1-D) mechanistic model of bedrock abrasion by total sediment load (Lamb et al., 2008b). Sediment supply has a strong influence on the fluvial incision of bedrock, enhancing incision by acting as a tool for abrasion and impeding incision by protecting the bed from grain impacts (Gilbert, 1877). Despite a recent proliferation of theoretical models translating this concept into mathematical formulations (Sklar and Dietrich, 2004; Turowski et al., 2007; Lamb et al., 2008b; Chatanantavet and Parker, 2009), field application has been mostly limited to inferring the relative dominance of tools and cover for specific fluvial systems (Turowski and Rickenmann, 2009; Cook et al., 2013; Ely et al., 2012). Prediction of fluvial incision rates or model testing at the reach scale has been rare (excepting Hobley et al., 2011; Beer and Turowski, 2015) owing to the difficulty in measuring high-resolution incision rates and key model parameters (e.g., sediment supply) in natural landscapes.

The undissected surface of Collier lava flow enables us to confidently calculate pre-incision topography and hence incision rate along the White Branch Creek (e.g., Ferrier et al., 2013), making this site uniquely suited to using mechanistic models of bedrock abrasion to model channel-forming discharges. Furthermore, the mixed bedrock-alluvial nature of the channel (Fig. 1) enables us to place constraints on long-term sediment supply, which is often impossible for bedrock channels. Our model results demonstrate that outburst floods are not necessary for channel incision, and they point to the importance of sediment availability for initiating and sustaining fluvial dissection.

**Volcanic and Glacial Setting**

The Cascade Range, which extends from northern California to southern British Columbia, is a volcanic arc associated with the subduction of the Juan de Fuca plate under the North American plate. In Oregon, the Cascades are divided into two physiographic provinces: the largely inactive, steep, and highly dissected Western Cascades, and the High Cascades, where numerous cinder cones, stratovolcanoes, and other vents have been active throughout the Quaternary (Jefferson et al., 2010). The high surface permeability of High Cascades deposits means that most precipitation in the High Cascades flows in the subsurface, feeding a sparse network of spring-dominated channels (Jefferson et al., 2010).

Most Cascade stratovolcanoes have been dissected by alpine glaciers. Moraine mapping points to multiple post–Last Glacial Maximum glacial highstands in the Oregon Cascades (Marcott et al., 2009), with the most recent occurring during the Little Ice Age, between A.D. 1750 and 1850 (Scott, 1977; O’Connor, 2001). Cascade glaciers have retreated substantially since the Little Ice Age, leading to several documented moraine-dam breaches in the past century (Hopson, 1960; O’Connor et al., 2001). Despite the overarching relationship between post-Little Ice Age retreat and moraine-dam breaches, these events can also be triggered by smaller-scale processes, such as wave overtopping (Clague and Evans, 2000), and interannual to decadal variability in glacier position (Anderson et al., 2014).

The Collier lava flow erupted from the Collier cone, a cinder cone on the western flank of North Sister in the High Cascades of Oregon, ~1600 yr
Rapid fluvial incision of a late Holocene lava flow

Flour”) in White Branch Creek as far downstream and that will be observed high loads of fine sediment (“glacial meltwater from the glacier discharged into the lake formed, bounded by the moraines and Collier cone. Between 1933 and 1960, the meltwater from this lake fed White Branch Creek, described by Hopson (1960, p. 4) as a “raging torrent” during peak summer daytime melt. Hopson also observed high loads of fine sediment (“glacial flour”) in White Branch Creek as far downstream as its confluence with the McKenzie River, a major regional drainage (~12 km downstream of the toe of Collier lava flow).

The terminal lake of Collier glacier drained catastrophically in 1942 (Hopson, 1960) producing a large sediment-laden flood with peak discharge as high as 500 m/s (O’Connor et al., 2001) that deposited sediment up to ~8 km downstream of the moraine-dam breach (Fig. 1). The breach is still visible as an ~5-m-deep incised notch between Collier cone and the lateral moraine (Fig. 1). The lake drained partially again sometime between 1954 and 1956, creating a new incisional notch to the north of the 1942 outburst, but this event did not have major downstream impacts (Hopson, 1960; O’Connor et al., 2001). There is no direct evidence that the channel we consider in this contribution was cut by these catastrophic events; deposits from the 1942 outburst terminate upstream of Collier lava flow (O’Connor et al., 2001). The lake level continued to decrease after these outbursts; after 1960, meltwater from the glacier discharged into the subsurface rather than into White Branch Creek.

METHODS

Measuring Channel Morphology and Incision

We used elevations from high-resolution airborne LiDAR (Deardorff and Cashman, 2012; NCALM, 2008) to map the morphology of White Branch Creek where it traverses Collier lava flow (Figs. 1 and 2). Because the channel lacks clear bankfull indicators (e.g., stripped vegetation; Fig. 3), we mapped the banks and thalweg of the gorge manually from LiDAR cross sections that were spaced every 5 m and oriented perpendicular to the channel. We labeled the lowest point in each cross section as the thalweg and marked prominent points of inflection on either side of the thalweg as banks (Fig. 2). To calculate incision in the gorges (Fig. 1), we mapped the location of the uneroded Collier lava flow surface along the length of White Branch Creek and used these elevations to interpolate a pre-incision surface (cf. Seidl et al., 1994; Ferrier et al., 2013). Local incision along the channel (Fig. 2) was then defined as the elevation of the pre-incision surface minus the elevation of the modern thalweg. We defined channel width for each cross section as the horizontal distance between the banks, delineated manually from LiDAR topography.

Alluvial Stratigraphy and Radiocarbon Sampling

To constrain the timing of channel-forming discharge events and discriminate between fluvial and debris-flow processes, we dug six soil pits in the alluvial deposits on Collier lava flow, three on the upper deposit and three on the lower deposit (Fig. 4), supplementing existing work by Deligne et al. (2013). We excavated each pit until we reached the lava flow surface or until the depth of the pit precluded continuing safely. We documented and measured the deposits (Fig. 5; Table DR1) and, where possible, collected charcoal for radiocarbon dating. Radiocarbon samples were processed and analyzed at the University of Arizona Accelerator Mass Spectrometer (AMS) laboratory following standard techniques. We used these data and stratigraphic relationships where available to estimate the age of the deposits using OxCal v.4.1.7 (Fig. 5; Bronk Ramsey, 2009).

FIELD EVIDENCE FOR INCISION MECHANISMS AND TIMING

White Branch Creek alternates between bedrock reaches and alluvial reaches as it traverses Collier lava flow (Figs. 1–3; Deligne et al., 2013). The boundaries between these reaches correspond with major breaks in the background slope and width of Collier lava flow (Fig. 1):

Figure 2. Channel morphology and topography: (A) long profile, (B) slope, (C) incision, and (D) valley width measured by differencing thalweg and interpolated pre-incision surface (see text for details). Raw elevation and valley width data (gray points in A and D) are smoothed (black circles and black line) to 100 m spacing (~10 channel widths) by fitting lines to the raw data upstream of each interpolated point. Large white circles in A correspond to channel cross sections in Figure 8. Gray boxes indicate extent of bedrock gorges. Note correspondence of maxima in slope and incision.

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Bedrock gorges occur where the flow is steep and narrow, while alluvial deposits are located in relatively flat areas. The bedrock gorges are significantly steeper (average gradient of 0.11) compared to the alluvial reaches (average gradient of 0.04; Fig. 4), mimicking the pattern of the background topography (Fig. 2). There is no evidence of significant sediment or water contributions from tributary channels (Fig. 1). The valley width of White Branch Creek does not vary systematically with distance downstream (a proxy for drainage area; see Eq. 3) in the alluvial or bedrock reaches (Fig. 2), despite past predictions and field observations that channel width tends to scale positively with drainage area (Leopold and Maddock, 1953; Montgomery and Gran, 2001).

Bedrock Gorges: Morphology, Pre-Incision Topography, and Sediment Provenance

Though rough and heterogeneous near the top where incision has resulted in bank failure, the lower part of the channel walls and bed of the bedrock reaches contain flutes and potholes (Fig. 3A), consistent with abrasion by sediment impacts (Whipple et al., 2000a; Sklar and Dietrich, 2004). At the 100 m scale (~10 channel widths), slope is strongly correlated with gorge depth (Fig. 6A), indicating that shear stress is the primary control on channel incision rather than longitudinal variations in rock hardness (contrast with Sklar and Dietrich, 2001). Channel slope is also correlated with the slope of the pre-incision lava flow surface (Fig. 6B). While the observed channel slopes are well in excess of typical fluvial values (up to 0.2; Montgomery, 2001), we saw no indication of debris-flow activity in surface morphology (i.e., levees, scarps) or stratigraphy (reverse-graded, poorly sorted, matrix-supported deposits; see next section).

Though the channel bed is generally bare, alluvial patches occur locally within the bedrock gorges. This alluvium contains clasts of Collier lava flow as well as abraded clasts of obsidian (Fig. 3B). The presence of obsidian strongly suggests that despite the absence of paleohydraulic indicators like those found upstream of the lava flow (e.g., gravel bars, overbank deposits; O’Connor et al., 2001), the discharge events responsible for carving the channel traversed the channel adjacent to the Obsidian Cliffs (Fig. 1), possibly en route from Collier glacier.

Alluvial Morphology and Sedimentology: Implications for Sediment Transport Rates and Mechanisms

Of the three alluvial deposits, we focused on the two downstream of the bedrock gorges, because they enable us to draw conclusions about the flux and caliber of sediment moving through the incisional reaches. The thick forest vegetation on these alluvial deposits is in stark contrast to the unaltered surface of Collier lava flow, which remains largely unvegetated (Deardorff and Cashman, 2012; Deligne et al., 2013). In these alluvial reaches, the channel of White Branch Creek is formed in the sediment of the deposits (e.g., Parker, 1978); i.e., there is no bedrock exposed in the bed. In addition to the main channel, there are several inactive channels on the alluvial deposits filled with organic debris, suggesting that the channel location in these reaches has not been fixed through time. When the active channel and exposed lava flow are excluded, the alluvial surfaces have a consistent slope of ~0.04 (Fig. 4C).

Mackin (1948) posited that over time, channel slope adjusts to provide sufficient power to both transport the sediment supplied to the channel and incise the channel bed at a rate matching the uplift rate. In incising bedrock
channels, this implies that the ability of the channel to transport sediment (known as sediment transport capacity) is in excess of long-term sediment supply (e.g., Sklar and Dietrich, 2006; Johnson et al., 2009), while in alluvial systems, such as the alluvial reaches of White Branch Creek, transport capacity is equal to or less than supply. Long-term transport conditions have also been demonstrated to control the surface slope (i.e., the average slope of the fan surface) of alluvial fans (Stock et al., 2008, and references therein). Given that the alluvial reaches of White Branch Creek are not incising, we can use the surface slope and channel geometry to calculate sediment transport capacity (see Model Formulation and Procedure), thereby setting a lower bound on the long-term rate of sediment supply in the bedrock gorges.

Sediment in the six alluvial soil pits (locations shown in Fig. 4) ranged from silt to large cobbles (Table DR1 [see footnote 1]). Each pit contained decimeter-scale normally graded packets (Fig. 5) with abrupt basal boundaries; buried organic material was present in the pits, but pedogenic alteration was minimal (Table DR1 [see footnote 1]; cf. Deligne et al., 2013).

Beds with coarse sediment (>4 mm) tended to be clast-supported; in some layers, clasts were imbricated (Table DR1 [see footnote 1]). The absence of reverse grading and matrix-supported deposits leads us to conclude that these sediments were not deposited by debris flows (Smith, 1986). Instead, we found evidence for emplacement by both hyperconcentrated flood flows (clast-supported deposits without preferred axis orientation, horizontal laminations in sand, normal grading) and normal traction flows (imbricated gravels, ripple cross-bedding; Smith, 1986), consistent with the findings of past work on Collier lava flow (Deligne et al., 2013). Because these depositional processes could conceivably occur with annual snowmelt, historical glacial meltwater, or outburst floods, our stratigraphic observations do not allow us to discriminate among these runoff mechanisms. Radiocarbon analyses of charcoal from two pits on the upper deposit (UF1 and UF2) and one on the lower deposit (LF2) yielded 95% confidence intervals of 580–397 cal yr B.P., 1345–1141 cal yr B.P., and 537–412 cal yr B.P., respectively (Fig. 7). These dates are consistent with (1) incision and associated deposition beginning shortly after lava flow emplacement 1600 yr ago, and (2) multiple events rather than a single catastrophic channel-forming flood (e.g., Lamb and Fonstad, 2010).

Hydrologic Regime

Field observations indicate two sources of discharge in White Branch Creek: snowmelt runoff and glacial outburst floods from moraine-dam breaches at Collier glacier. During June 2008, we observed shallow discharge in both the bedrock gorges and alluvial reaches, which, given that the snowpack of 2007–2008 was not exceptional (SNOTEL, 2016), we infer occurs on an approximately annual basis. Though snowmelt discharge here appears minor, past
work in glacial landscapes shows that meltwater events can transport a significant amount of sediment (Sanders et al., 2012), and, as noted already, White Branch Creek was a significant pathway for glacial meltwater between 1933 and 1960.

As noted earlier, there is also historical documentation of moraine-dam breaches at Collier glacier (Hopson, 1960). Though it is does not appear that these recorded events reached the uppermost alluvial deposit (O’Connor et al., 2001), the presence of obsidian clasts and other externally sourced sediment on the lava flow leads us to conclude that similar events may have traveled farther downstream (i.e., onto the lava flow) in the past. As noted already, these glacially derived floods can occur even in the absence of a regional climatic event (e.g., the Little Ice Age) due to local variations in glacier position and impoundment stability (Anderson et al., 2014).

**MODELING PHILOSOPHY: ASSESSING THE DRIVING FORCES OF INCISION**

Despite our ability to describe much of this system through field observation, LiDAR topography, and alluvial stratigraphy, it is not clear from the evidence described in the previous section how the two discharge regimes combine to provide geomorphic power to the channel. Could the channel have been carved solely by glacial outburst floods, or does annual snowmelt discharge play a role? To answer these questions, we set out to model the incision of Collier lava flow numerically, using the channel morphology measured from LiDAR-derived topography as model input and constraining other model parameters with our field observations and inundation models.

In selecting an appropriate model, we began by considering the simplest formulation that could explain the discharge events driving the formation of White Branch Creek. The close correlation between channel slope and local incision (Fig. 8) suggests that the stream power law of bedrock erosion (Howard, 1994) may provide an adequate description of this system, where erosion rate $E$ is a function of channel slope $S$, drainage area $A$ (a proxy for discharge), and positive constants $K$, $m$, and $n$:

$$E = KA^mS^n.$$  

(1)

We fit the stream power model to the incision reaches of White Branch Creek by rewriting Equation 1 as:

$$\log E = \log K + m \log A + n \log S.$$  

(2)

and using multiple linear regression to find the best-fit values of $K$, $m$, and $n$ (Seidl et al., 1994). The high surface roughness of Collier and Sims Butte lava flows (Fig. 1B) confounds the flow-routing algorithms typically used to calculate drainage area, so we instead use Hack’s law (Hack, 1957) with an $h$ exponent of 0.57 to convert distance downstream, $x$, to drainage area:

$$A = x^{1/h}.$$  

(3)

We fit Equation 2 to (1) all incisional reaches, (2) only the upper gorge, and (3) only the lower gorge (Table 1). For all three cases, there is a significant positive correlation between slope and incision (Table 1), as expected. Surprisingly, the fit values of $m$ were significantly negative, suggesting that incision decreases with downstream distance.

The stream power model is useful for tectonic geomorphology, where steady-state channel morphology can be taken as a proxy for uplift rate (Wobus et al., 2006; Kirby and Whipple, 2012), or in comparisons of channels incising into different material (Sklar and Dietrich, 2001) or under different climate conditions (Ferrier et al., 2013). For the case of White Branch Creek, the negative correlation between drainage area and incision has interesting hydrologic implications (see Discussion). However, these parameter values do not yield mechanistic insight into the erosive capacity of the individual discharge events that likely occurred in this channel.

Given this limitation, and the fact that our motivation for this study was investigation of sediment-driven mechanisms of erosion, we instead used a formulation that describes the physical process of fluvial erosion of bedrock via sediment abrasion (Lamb et al., 2008b). Bedrock incision occurs via two primary processes: plucking, whereby jointed blocks are removed from the bed by flowing water (Whipple et al., 2000a; Chatanantavet and Parker, 2009; Lamb and Dietrich, 2009), and abrasion, whereby sediment grains impacting the bed gradually erode it via mechanical wear (e.g., Sklar and Dietrich, 2004). Plucking tends to dominate in highly fractured bedrock and produces large blocks, while abrasion dominates in massive, intact bedrock and forms potholes, flutes, and other sculpted features (Whipple et al., 2000a).

Despite past observations of plucking in basaltic bedrock, our field observations suggest that abrasion is the dominant process of incision at White Branch Creek. The channel has potholes, smooth walls, and flutes (Fig. 3A; Wohl and Merritt, 2001), and we observed limited...
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Large blocks or cavities that might indicate a plucking-dominated channel (e.g., Whipple et al., 2000b; Lamb and Fonstad, 2010). The channel bed is primarily established in the coherent inner core of the lava flow, rather than the rubbly surface layer, which was presumably rapidly removed early in the incision process (e.g., Whipple et al., 2000b).

MODEL FORMULATION AND PROCEDURE

Assumptions and Parameter Constraints

All sediment-dependent models of bedrock incision (Sklar and Dietrich, 2004; Turowski et al., 2007; Lamb et al., 2008b; Chatanantavet and Parker, 2009) require bed shear stress, grain size, and sediment supply as inputs. Our field observations allowed us to put bounds on the range of these quantities, but we lacked sufficient information to constrain the most likely parameter combinations or to define the shape of the input parameter distributions. In acknowledgment of this uncertainty, we used a Monte Carlo algorithm in which we selected a different set of these three parameters (channel discharge [and shear stress], grain size, and sediment supply) for each model run from a uniform distribution of possible values, the bounds of which were set as described next.

Channel Hydraulics

Converting discharge to bed shear stress requires quantifying flow hydraulics. The hydraulics of steep mountain channels are sufficiently complex that there is no accepted formulation for bed shear stress in these environments (cf. Wohl, 2000). Instead, we followed previous authors (e.g., Sklar and Dietrich, 2004) in using the depth-slope product to calculate bed shear stress for White Branch Creek:

\[ \tau_s = \rho_w g H S, \]  

where \( \rho_w \) is the density of water, \( g \) is the acceleration of gravity, and \( H \) is the hydraulic radius. Because White Branch Creek in ungauged, and neither of the historic outburst floods from Collier Glacier was observed in our study reach, we calculated the hydraulic radius using the Hydraulic Engineering Center’s River Analysis System (HEC-RAS). This software uses the energy equation to calculate inundation at discrete channel cross sections for a given discharge. Our input topographic data for these calculations were 50-m-long channel cross sections derived from LiDAR topography, spaced every 5 m and oriented perpendicular to the channel thalweg.

Due to the high channel slopes of White Branch Creek (see discussion in Field Evidence section), the energy equation predicts supercritical flow conditions for the length of the channel (Fig. DRI [see footnote 1]). Prolonged supercritical flow is unreasonable in hydraulically rough channels like White Branch Creek, which have high rates of energy dissipation (Montgomery and Buffington, 1997). Hence, we restricted the flow conditions in HEC-RAS to subcritical or critical flow (Grant 1997; Comiti et al., 2009).

The energy equation procedure described here relies on a description of hydraulic roughness, or the tendency of energy to dissipate. To describe roughness, HEC-RAS uses the empirical relationship in Manning’s equation, where average channel velocity \( U \) depends on slope \( S \), hydraulic radius \( R_h \), and Manning’s \( n \), a proxy for roughness:

\[ U = \frac{R_h^{2/3} S^{1/2}}{n}. \]  

The value of Manning’s \( n \) for natural channels depends on many factors, including grain size relative to channel width, the degree of channel obstruction, and the presence and type of vegetation (Phillips and Tadayon, 2006). Limerinos (1970) presented an approach for calculating Manning’s \( n \) based on hydraulic radius and the \( D_{84} \) (84th percentile of the grain size distribution) of the bed material:

\[ n = \frac{0.8204 R_h^{1/6}}{1.16 + 2.0 \log_{10} \left( \frac{R_h}{D_{84}} \right)}. \]  

As shown in the LiDAR cross sections and HEC-RAS inundation models (Figs. 2 and 8), the hydraulic radius of the channel varies between 1 and 6 m, depending on location and discharge magnitude. Based on field observations, the \( D_{84} \) of the bed material is between 0.01 m and 0.5 m (including in-channel boulders). We used Equation 6 to calculate Manning’s \( n \) for the roughest (minimum \( R_h \), maximum \( D_{84} \)) and

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<th>Table 1. Stream-power parameters fit to channel data from incisional reaches</th>
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Note: \( K \)—best-fit stream power constant, \( m \)—drainage area exponent, \( n \)—slope exponent. \( R² \)—variance explained by regression. \( SE \) columns show standard error for all fit parameters.
smoothest (maximum $R$, minimum $D_{50}$) cases, yielding Manning’s $n$ values between 0.02 and 0.06. This range of Manning’s $n$ values has a surprisingly modest effect on channel depth and width (Fig. DR2 [see footnote 1]); model results are presented for Manning’s $n = 0.04$ (see Fig. DR3 [see footnote 1] for effect of Manning’s $n$ on model output).

**Representative event discharges.** A primary goal of this work was to model discharges ranging from annual snowmelt flows to large outburst floods. To estimate the discharge associated with annual snowmelt, we return to the field observation of 0.5 m flow depth in a 2.0-m-wide channel reach during the snowmelt of 2008 (Fig. 3C). If the full range of grain sizes observed in the channel is transported during annual snowmelt events, then at a minimum, the flow must be capable of moving the median grain size ($D_{50}$) of bed material (0.03 m). Sediment motion begins when the Shields number $\tau_s$ exceeds some critical value, which we took here to be 0.045 (Sklar and Dietrich, 2004). To calculate bed shear stress, we used the relationship

$$\tau_s = \tau_c (\rho_s - \rho_w) g D_{50}, \quad (7)$$

where $\rho_s$ is the density of sediment, $\rho_w$ is the density of water, and $g$ is the acceleration of gravity. Following the approach of Lamb and Fonstad (2010), we used a modified Manning-Strickler relationship (Parker, 1991) to calculate the discharge:

$$Q = UA = 8.1A \left( \frac{\tau_s}{\rho_w} \right)^{1/2} \left( \frac{h}{k_s} \right), \quad (8)$$

where $U$ is the average flow velocity, $A$ is the cross-sectional area of the channel, $h$ is channel depth, and $k_s$ is the roughness length scale. This approach is similar to Manning’s law (Eq. 5), with the difference that roughness is quantified as a length scale rather than as a bulk parameter. We varied $k_s$ between 0.08 m (2 × $D_{50}$; Parker, 1991) and 0.5 m (the approximate radius of boulders in the channel) to capture the considerable longitudinal variability in channel roughness. For the flow depth and cross-sectional area observed in the field, we calculated a representative snowmelt discharge between 1.1 and 1.6 m$^3$/s. As an alternate approach, we used Equation 5 to calculate channel velocity $U$ for a Manning’s $n$ between 0.02 and 0.06 where channel geometry ($S$ and $R_f$) are measured from LiDAR (Fig. 2) is 0.15. This approach yields snowmelt discharges ranging from 3.1 to 9.3 m$^3$/s.

Hydraulic reconstructions of historical outburst floods from Collier glacier show peak discharges of 40 m$^3$/s at the downstream extent of the flood. To allow for the possibility that prehistoric floods exceeded the magnitude of historical floods, we also calculated maximum outburst flood discharge using the “brimful” assumption (Baker and Milton, 1974), which is based on the idea that the flood discharge filled the deepest part of the canyon. Under steady, uniform, subcritical flow conditions, our HEC-RAS runs show that the deepest channel cross section in the gorge (Fig. 8A) was filled by 400 m$^3$/s. This value is similar to the estimate of peak discharge at the breach site for the 1942 outburst event (500 m$^3$/s; O’Connor et al., 2001). **Grain size.** Sediment in the modern bedrock channel and on the alluvial plains ranges in size from silt to coarse cobbles. Due to this breadth of grain sizes, we erred on the side of inclusion for our model inputs, selecting median grain size diameters between 6.3 × 10$^5$ and 1.5 m. **Sediment supply.** As discussed already, in nonincising fluvial settings, landforms are created by the balance between sediment transport capacity and sediment supply (Stock et al., 2008; Johnson et al., 2009). With this in mind, we used the surface slope of the alluvial deposits and the average geometry of the modern alluvial channel to calculate bed-load transport capacity $q_{bc}$ in the alluvial reaches. For consistency with the literature (Sklar and Dietrich, 2004; Lamb et al., 2008b; Scheingross et al., 2014), we used the Fernandez Luque and Van Beek (1976) relation for bed-load transport capacity, which states that sediment transport capacity is a power-law function of bed shear stress:

$$q_{bc} = 5.7 (R g D)^{3/2} (\tau_s - \tau_c)^{3/2}, \quad (9)$$

where $R$ is the specific density of sediment, and $D$ is the average diameter of sediment grains. For each model run, we used the selected values of discharge and grain size, and the average width of the modern alluvial channel to calculate bed shear stress in the alluvial reaches (Eq. 4). The bed-load transport capacity calculated by Equation 9 was then used as the lower bound of the sediment supply parameter space. We set the upper bound at 20 m$^3$/s (width-averaged), which is sufficiently large to cover the bed at the highest modeled discharge (Fig. 9G). In using modern channel geometry to calculate bed-load transport capacity, we assumed that channel shape reflects long-term average flow conditions (e.g., Parker, 1978; see Discussion).
Mechanistic Modeling of Bedrock Abrasion

Mechanistic models of fluvial abrasion by sediment impacts state that erosion rate \( E \) is a product of the average volume of sediment removed per impact, \( V \), the rate of impacts per unit bed area, \( I \), and the fraction of the bed that is exposed to abrasion, \( F \) (Sklar and Dietrich, 2004; Turowski et al., 2007; Lamb et al., 2008b):

\[
E = VIF. \quad (10)
\]

The first version of these models (Sklar and Dietrich, 2004) considered only impacts from sediment traveling as saltating bed load and used the hop length and kinetic energy of sediment grains to calculate \( I \) and \( V \). As transport stage increases toward the threshold of suspension (\( \tau \approx 20 \)), the hop lengths become infinitely long, making \( I \) (and, by extension, \( E \)) approach zero. Lamb et al. (2008b) recast this model in terms of total load, recognizing that due to turbulent fluctuations, even sediment traveling in suspension comes in contact with the bed and may be important to fluvial abrasion. Their model described particle velocity (and hence impact energy) as a sum of the particle motion due to gravitational settling and turbulent advection, using a Gaussian approximation for turbulent fluctuations and a modified Rouse equation to calculate the relationship between sediment concentration and bulk flow parameters (Lamb et al., 2008b).

For most bedrock systems, the contribution of suspended load to erosion is relatively small compared to bed load, but for rivers with large discharges and/or small grain size (i.e., large transport stage), the difference between bedload-only models and the total load model is not negligible (Scheingross et al., 2014). Specifically, the bedload-only model of Sklar and Dietrich (2004) would tend to predict zero erosion in these systems, while the total load model (Lamb et al., 2008b) predicts nonzero erosion. Due to the high discharge of outburst floods (O’Connor et al., 2001) and the relatively small grain sizes we observed in the alluvial deposits, we used the total load formulation (Lamb et al., 2008b) to model instantaneous incision rates in White Branch Creek.

Governing Equations

To calculate erosion rate \( E \) with the total load model, we used Equation 36 of Lamb et al. (2008b), a derivation of Equation 10 that uses a combination of empirical relationships and physical arguments to calculate \( V \), \( I \), and \( F \) (Lamb et al., 2008b). This equation states that \( E \) is a function of material properties, sediment flux, the partition of sediment between suspension and bed load, and channel flow:

\[
E = \frac{A_1 \rho_Y}{k \sigma_I} \left( \frac{q_{v_{\rm eff}}}{U H b} \right)^3 \left( 1 - \frac{q_{v}}{q_{v_b}} \right), \quad (11)
\]

where \( A_1 \) is the cross-sectional area of a sediment grain; \( k \) is an empirical erodibility constant; \( \sigma_I \) is the tensile strength of the rock; \( \rho_Y \) is the density of sediment; \( Y \) is Young’s modulus; \( q \) is the volumetric sediment supply per unit width; \( w_{v_{\rm eff}} \) is the effective impact velocity of sediment (derived from the normal component of settling velocity); \( U \) is the streamwise flow velocity; \( H \) is the flow depth; \( I \) is an integral relating suspended sediment flux to sediment concentration, flow depth, and velocity; \( U_i \) is the streamwise bed-load velocity; \( H_b \) is the thickness of the bed-load layer; \( q_{v} \) is the volumetric bed-load flux per unit channel width; and \( q_{v_b} \) is the volumetric bed-load transport capacity per unit channel width (Lamb et al., 2008b).

The first group of terms on the right-hand side of Equation 11 consists of material parameters of the sediment and the substrate, the second group of terms describes sediment concentration and impact velocity, and the final group of terms in parentheses is the percent of bedrock exposed to erosion. To convert the alluvial bedload flux \( q_{v_b} \) (calculated from the alluvial reach morphology) into total sediment supply \( q \), we used an empirical description of bed-load layer height and velocity, the vertical structure of sediment concentration, and depth-average flow parameters (Eq. 20 of Lamb et al., 2008b) and solved for \( q \):

\[
q = q_{v_b} \left( \frac{U H b + U_i H_i}{U_i H_i} \right) \quad (12)
\]

where the \( a \) subscript indicates bed-load parameters for the alluvial reaches.

Modeling Procedure

We predicted instantaneous incision rates along White Branch Creek using the 1-D longitudinal channel profile (sampled at a geomorphically relevant interval of ~10 channel widths [100 m]) and a constant channel width. Values for all other parameters are given in Table 2. We used the following procedure to calculate incision rates for a range of discharges, performing 500,000 iterations for each discharge value:

1. Select value of grain size from uniform distribution.
2. Calculate flow depth and width in the gorge using HEC-RAS.
3. Calculate minimum sediment supply based on shear stress (Eqs. 4 and 9), grain size, slope of alluvium (Fig. 4C), and alluvial channel geometry; this sets the lower bound of the uniform distribution of sediment supply.
4. Select value of sediment supply from uniform distribution; convert to total load sediment supply (Eq. 12).
5. Calculate instantaneous incision rate at each point along the gorge (Eq. 11).
6. Calculate median erosion rate for given parameter set.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model output</td>
<td>Incision rate (L/T)</td>
</tr>
<tr>
<td>Model inputs</td>
<td>Total load sediment supply per unit channel width (L^2T^{-1}L^{-1})</td>
</tr>
<tr>
<td></td>
<td>Grain size (L)</td>
</tr>
<tr>
<td></td>
<td>Water discharge (L^2T^{-2})</td>
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<td></td>
<td>Channel slope (L/L)</td>
</tr>
<tr>
<td></td>
<td>Width (L)</td>
</tr>
<tr>
<td>Calculated values</td>
<td>Volume eroded per impact (L^2T^{-1})</td>
</tr>
<tr>
<td></td>
<td>Impact rate (T^{-1})</td>
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<tr>
<td></td>
<td>Depth-averaged streamwise water velocity (LT^{-1})</td>
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<tr>
<td></td>
<td>Depth-averaged streamwise bed-load velocity (LT^{-1})</td>
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<tr>
<td></td>
<td>Water depth (L)</td>
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<tr>
<td></td>
<td>Depth of near-bed sediment layer (L)</td>
</tr>
<tr>
<td></td>
<td>Integral relating the flux of suspended sediment to sediment concentration, depth, and velocity (dimensionless)</td>
</tr>
<tr>
<td></td>
<td>Bed-load supply per unit width (L^2T^{-1})</td>
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<td></td>
<td>Bed-load transport capacity per unit width (L^2T^{-1})</td>
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<td></td>
<td>Effective impact velocity (LT^{-1})</td>
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<tr>
<td></td>
<td>Cross-sectional area of sediment grain (L^2)</td>
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<td></td>
<td>Shields stress (dimensionless)</td>
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<tr>
<td></td>
<td>Critical Shields stress for incipient motion (dimensionless)</td>
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<tr>
<td></td>
<td>Density of sediment (ML^{-3})</td>
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<td></td>
<td>Density of water (ML^{-3})</td>
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<td></td>
<td>Young’s modulus (ML^{-1}T^{-2})</td>
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<tr>
<td></td>
<td>Empirical bedrock erodibility coefficient (dimensionless)</td>
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<tr>
<td></td>
<td>Rock tensile strength (ML^{-1}T^{-2})</td>
</tr>
<tr>
<td></td>
<td>Specific density of sediment (dimensionless)</td>
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<tr>
<td></td>
<td>Acceleration of gravity (LT^{-2})</td>
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</tbody>
</table>

Note: L—length, M—mass, T—time.
EROSION BY SNOWMELT AND OUTBURST FLOODS

We calculated instantaneous erosion rates for discharges from 1 to 400 m$^3$/s (Figs. 9 and 10). For each discharge, we report the median spatially averaged nonzero incision rate for all model runs (Fig. 10). Given the lack of constraints on the time evolution of channel geometry, we did not consider the spatial pattern of erosion. The instantaneous erosion rate (Figs. 9D and 10) is a function of three main factors: effective sediment impact velocity (Fig. 9A), the fraction of the bed exposed (Fig. 9B), and near-bed sediment concentration (Fig. 9C; Lamb et al., 2008b). All three quantities depend on transport stage, and near-bed sediment concentration and relative sediment supply also depend on sediment supply (Lamb et al., 2008b). For a given model discharge, flow depth and channel slope are constant, meaning that transport stage is a function of grain size (Fig. 9).

Figure 9 shows the effect of grain size on effective impact velocity (Fig. 9A), near-bed sediment concentration (Fig. 9B), relative sediment supply (Fig. 9C), and erosion rate (Fig. 9D) for two modeled discharges: 10 m$^3$/s, the highest representative snowmelt discharge calculated, and 400 m$^3$/s, a representative magnitude for an outburst flood (see earlier herein for details on representative discharges). Unsurprisingly, erosion rates are lower for 10 m$^3$/s than 400 m$^3$/s because of lower average shear stress. There are also fewer combinations of parameters that result in nonzero erosion for 10 m$^3$/s than for 400 m$^3$/s (3681 out of 500,000 model runs compared to 118,761 of 500,000); the lower shear stress means that the flow is often below the critical shear stress for mobilization (transport stage <1), the flow cannot transport the full load of sediment (fraction of bed exposed = 0), or the grains are larger than the depth of the flow ($D > H$).

An examination of the modeled components of erosion rate (Figs. 9A–9C) for 400 m$^3$/s reveals how grain size and sediment supply control erosion via abrasion. The spread in values for a given grain size is due to different values of sediment supply. Effective impact velocity generally increases with grain size because larger grains have higher settling velocities (Dietrich, 1982). For the 10 m$^3$/s case, effective impact velocity begins to decrease around $D = 0.8$ m ($D$ is grain size), when the distance that particles fall through the water column ($D_{\text{fall}}$ of Lamb et al., 2008b) is not large enough for the particles to reach terminal settling velocity. Near-bed sediment concentration depends primarily on sediment supply, but it generally decreases with grain size; model runs with high grain size and high near-bed sediment concentrations result in no erosion because the bed is covered by sediment. The patterns of these three quantities combine to produce the pattern of erosion rate with grain size shown in Figure 9D. The minimum value in the upper bound of erosion rate at ~0.1 m grain size corresponds to the increase in the fraction of exposed bedrock, as well as a flattening of the effective impact velocity curve (Fig. 9A).

Modeled instantaneous incision rates in the gorge reaches of White Branch Creek range from $0.7 \pm 0.4$ m/yr for annual snowmelt (1 m$^3$/s) to $42 \pm 47$ m/yr (Fig. 10A). The uncertainty in these values captures the variance due to different grain size and sediment supply parameter combinations; we propagated this uncertainty in the following calculations via Gaussian error propagation. By dividing total average incision in the gorges (~5 m) by the incision rate for each discharge, we calculated the implied time of active incision (Fig. 10B) using the nonzero erosion rates. For the lower end of our discharge range (annual snowmelt; 1 m$^3$/s), 6.5 ± 3.9 yr of continuous incision would be required to form the gorges of White Branch Creek; large events (400 m$^3$/s) would need 0.12 ± 0.13 yr. Event durations for both outburst floods and snowmelt are likely on the order of hours to days. If we assume a 1 d duration for both types of events, we would need 2300 ± 1400 d of snowmelt or 44 ± 47 outburst floods to accomplish the incision we observed. Given the lava flow is 1600 yr old, this requires only 1.4 ± 1.1 days per year of snowmelt. Our model results indicate that incision could be accomplished via annual flows alone, given the right sediment supply and grain size, and that we do not need large outburst floods to cause fluvial incision. Furthermore, between 1933 and 1960, when White Branch Creek drained the meltwater from Collier glacier, noncatastrophic discharge probably occurred throughout the summer months at higher discharges and with higher sediment loads than modern snowmelt (the “raging torrent” of Hopson, 1960).

DISCUSSION

Channel Width Scaling, Stream Power, and Implications of the Drainage Area Exponent

In alluvial and bedrock channels worldwide, channel width has been shown to scale negatively with slope (Finnegan et al., 2005) and positively with drainage area (Leopold and Maddock, 1953; Montgomery and Gran, 2001). Our measurements of valley width in the bedrock reaches of the White Branch do not display this pattern; in fact, in the upper gorge, valley width is roughly positively correlated with channel slope. One explanation might be the effect of incision on the side slopes: Although the channel is cut into the coherent core of the lava flow, much of the channel bank is composed of the rubbly surface layer. As the channel in-
Rapid fluvial incision of a late Holocene lava flow

Hydraulics, Sediment Transport, and Application of Incision Models to Steep Channels

We made several assumptions about the relationships among channel morphology, flow, shear stress, and sediment flux. By using the depth-slope product (Eq. 4) to calculate bed shear stress, we likely overpredicted the true bed shear stress available for sediment transport (e.g., Yager et al., 2007). Furthermore, other factors, such as grain-grain interactions that affect sediment mobility, result in overprediction of bed-load flux in steep mountain streams using most sediment transport equations (e.g., Eq. 9; Yager et al., 2007). This overprediction could result in a nonzero model erosion rate for parameter combinations where sediment supply actually exceeds the transport capacity. Hence, although our model predicts nonzero erosion for characteristic snowmelt runoff given certain sediment conditions, depending on the magnitude of our overprediction, some of these events may be below the critical shear stress for sediment motion or have insufficient transport capacity to convey the input sediment supply.

In using the morphology of the alluvial reaches to constrain minimum long-term sediment supply, we assumed that (1) these features reflect a long-term balance between supply and transport, and (2) that all discharge events contribute to the deposition of sediment on the alluvial plain. Larger discharge events are more likely than annual snowmelt flows to produce the overbank events (Fig. 8) and channel avulsions, however, and may therefore be the primary control on the morphology of the alluvial sections. If these features are not equally created by all discharges, it is also possible that the surface morphology is not at steady state, as assumed in our model.

The known heterogeneity of sediment transport and shear stress distributions in steep mountain channels would seem to argue against applying a mechanistic model that explicitly includes sediment transport relationships that are poorly constrained or inappropriate for these systems. However, the assumptions we make here regarding the relationship between discharge and bed shear stress are also inherent in the widely used stream power model (Eq. 1); sediment effects on incision are simply subsumed in the fit parameters K, m, and n. In using a sediment-dependent model for the White Branch Creek, our work shows the variability in incision rate that can occur due to differences in grain size and sediment supply under constant-flow conditions. For example, our results indicate that despite low discharge relative to glacial outburst floods, annual snowmelt flows are theoretically capable of rapidly carving a channel in a steep landscape. Similarly, the modeled incision rates for 40 m3/s, a representative outburst flood, span an order of magnitude over a reasonable range of grain size and sediment supply values, despite the same discharge conditions. We can also quantify the grain size and sediment supply responsible for our modeled incision rates (Fig. DR4 [see footnote 1]) and observe that maximum incision for a given discharge occurs for the largest grains transported for each discharge (Fig. DR4 [see footnote 1]). For smaller discharges, the maximum erosion is generated by low sediment supply values, indicating that another advantage of using mechanistic models lies in the ability to easily accommodate and test alternative descriptions of model components, such as the bedrock cover term (Turowski et al., 2007; Hobley et al., 2011).

Channel Initiation

Given that both modern snowmelt runoff and glacial meltwater are capable of cutting into bedrock, is it possible that outburst floods have no role at all in incision? The modeling described herein focused on the incision of the coherent core of the lava flow, implicitly assuming that a nascent channel form already existed. The work described in this paper does not shed light on how this proto-channel initiated, but we speculate that the high, fine sediment loads associated with glacial meltwater flows may have plugged the permeable rubbly surface, enabling surface flow to converge and facilitate incision. Alternatively, the channel may have initiated via plucking of the rubbly surface blocks (e.g., Lamb and Fonstad, 2010), converging to a single channel once the coherent core was exposed. Using Equation 7 and the approximate diameter of the rubbly surface blocks (1 m), we calculated that 728 Pa of shear stress would be necessary to initiate block motion. Given that the gradient of the pre-incision surface is up to 0.2, and using the depth slope product to solve for hydraulic radius, this would only require a flow with a hydraulic radius of 0.37 m (Eq. 2); our HEC-RAS results for 10 m/s have a median hydraulic radius of 0.43 m. Hence, though outburst floods may have affected our study reaches, these extreme events are not necessary to initiate channel incision or explain the magnitude of erosion we observed in the field.

Evolution of Temperate Volcanic Landscapes

Large floods are widely recognized as geomorphic drivers in jointed bedrock, where canyon carving can happen within hours via detachment of large blocks from the channel bed. Here, we demonstrate that fluvial incision of a blocky lava flow may be initiated and sustained by moderate snowmelt and glacial meltwater flows. We speculate that it is not large discharge events that distinguish the geomorphic evolution of the Collier lava flow from lava flows of the same age, but the abundance of tools provided by the moraines of Collier glacier. The presence of sediment in this system is crucial for enabling flows to erode the bed; without tools, the Collier lava flow would likely remain undissected and barren like many Quaternary lava flows in the High Cascades. Though our study site may be exceptional in its combination of alpine glaciation and volcanic activity, our findings have global implications, suggesting that large sediment sources such as moraines or landslide deposits may accelerate fluvial dissection by providing tools to otherwise sediment-starved systems (Wu et al., 2016).

CONCLUSIONS

We exploited the excellent topographic and temporal constraints of the Collier lava flow in the Oregon High Cascades to investigate the early stages of bedrock incision in young volcanic landscapes. LiDAR analysis and field observations of the morphology and stratigraphy of this mixed bedrock-alluvial system show that: (1) the channel has been carved by multiple discharge events, and (2) the pre-incision lava flow topography is the primary control on channel morphology at the reach and subreach scale, dictating both the location of the bedrock-alluvial transitions and channel slope within the two bedrock gorges. Based on field evidence...
that the channel conveys both annual snowmelt discharge and large, infrequent outbursts from Collier glacier, we used a numerical model of bedrock abrasion via total load (Lamb et al., 2008b) to calculate instantaneous incision rates for a range of sediment supply and grain-size scenarios. Our model results show that due to high channel slopes and abundant sediment in this system, even moderate discharges can drive appreciable rates of bedrock abrasion. Furthermore, given a historically active glacial meltwater system, catastrophic flows are not necessary to initiate channel formation. The ability of moderate-discharge events to incise these landscapes is contingent on the presence of abundant sediment from glacial moraines, suggesting that sediment availability may be a limiting factor in the onset of fluvial incision in young volcanic landscapes worldwide.

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