Climatic control of denudation in the deglaciated landscape of the Washington Cascades

This document 1) explains the chemical procedure of cosmogenic $^{10}$Be extraction, 2) calculates short-term, long-term, and Quaternary denudation rates including correction of cosmogenic $^{10}$Be production rates, 3) discusses potential biases in $^{10}$Be denudation rates from previous glaciations, 4) shows topographic metrics (channel steepness index for river incision, failure index for shallow landslides), and 5) explains that the increase of short-term denudation rates is not derived from tectonic rate changes, but from geomorphic process changes.

First, total denudation rates were measured from cosmogenic $^{10}$Be concentrations in thirteen detrital river sand samples across the Washington Cascades (Fig. 1b). Sampled basins are located across the precipitation gradient with minimum lithologic variation, which intends to isolate the climatic impact on the relations among denudation, topography, and climate within this landscape. The basins are underlain mostly by granites, granodiorites, and gneiss (although schist, shale, and metasedimentary rocks constitute a lesser component of rocks exposed in the landscape). The 250-500 and 500-710 µm size fractions were extracted from bulk river sand samples, and ~ 100 g of quartz was isolated using magnetic separation followed by progressive HF/HNO$_3$ leaches. Next, the quartz was digested with ~ 300 µg of Be carrier, and Be was extracted and oxidized to produce BeO$^3$. The $^{10}$Be/$^{9}$Be ratios were measured at Lawrence Livermore National Laboratory’s Center for Accelerator Mass Spectrometry facility using the calibration in Nishiizumi et al. The $^{10}$Be/$^{9}$Be ratio for blank samples was $0.63 \pm 0.16 \times 10^{-14}$. The blank was subtracted and its uncertainty was propagated into the concentrations. The $^{10}$Be concentrations corrected with analytical and blanks errors are presented in Supplementary Table S1.

Second, the Be concentrations were converted to basin-wide averaged denudation rates using the mean catchment elevation and time-dependent corrections of Lal and Stone implemented in version 2.2 of the CRONUS-Earth $^{10}$Be–$^{26}$Al age calculator (Information used for denudation calculation is presented in Supplementary Table S1). Since the CRONUS-Earth calculator does not account for the spatial variability of production rates that result from both latitudinal changes in basin locations as well as changing elevations throughout the basin, we calculated basin-averaged scaling factors of production rates considering the non-linearity of topography and shielding of cosmic radiation by topography and snow cover. We calculated three pixelwise scaling factors using 100-m-resolution DEM based according to scaling scheme of Lal and Stone. First, we considered the scaling factor for the nonlinearity of production rate with elevations, which is calculated from the ratio between production rate based on mean elevation and production rate based on real topography. Second, since the study area has significant relief and high altitude, we took into account the shielding from topographic obstacles. Topographic shielding was calculated for each pixel based on the equation below:

$$S_{\text{topo}} = 1 - \frac{1}{36} \sum_{i=1}^{n} \Delta \varphi \sin^{-1} \theta_i$$

Where $S_{\text{topo}}$ is the shielding factor for topography, $n$ is numbers of segments with an
azimuthal angle of the incoming radiation $\Delta \phi$, and with an inclination angle from the horizontal $\theta$, and an experimental constant $m \sim 2.3$. The 360° shielding factor was calculated for each DEM pixel with 10° steps for the azimuth angle. Third, a shielding factor for snow cover was calculated from monthly snow depth data which were averaged over 1961-1990 (Climate Atlas of the United States). Using the relationship between elevation and monthly snow depth inferred from the 29 snow stations near our study area, we interpolated the spatial distribution of snow depth over the study area. Assuming a maximum snow density of 0.3 g/cm³, the monthly snow shielding factors were estimated for each pixel. The averaged snow shielding factor was calculated from the average of monthly shielding factors:

$$S_{\text{snow}} = \frac{1}{12} \sum_{i=1}^{12} e^{-\left(\frac{z_{\text{snow},i}}{\rho_{\text{snow},i}}/\Lambda\right)}$$

Where $S_{\text{snow}}$ is the shielding factor from snow cover, $\rho$ is the maximum density of 0.3 g/cm³ for snow, and $z$ is the snow thickness (cm) and $\Lambda$ is the attenuation length (160 g/cm²). The scaling factor for non-linearity of topography, shielding for topography, and shielding for snow cover were all multiplied for each pixel, and averaged over all the points within the basin. The basin-averaged scaling factor was used in CRONUS calculator. The scaling factors for each basin are presented in the Table S1.

Long-term denudation rates were calculated from apatite (U-Th)/He (A-He) analyses (P. W. Reiners, personal communication) using a one-dimensional, steady state denudation and thermal model. These ages were interpolated throughout the study area using a kriging algorithm with a spherical model variogram to calculate basin averaged exhumation rates. We follow the authors’ interpretation of the long-term denudation rates derived from A-He ages as reflecting the long-term rock uplift rates (the rate at which rock moves upward relative to a stable datum such as sea level). The point measured range of long-term denudation rates are 0.01-0.33 mm/yr, but the averaged long-term denudation rates of our sample basin range from 0.03-0.22 mm/yr. Our denudation rates are on average 4 times higher than long-term denudation rates, 2 times higher than highest rates. We present the long-term denudation rates with the range of minimum to maximum long-term rates in error bars, which will show the full range of possible long-term rates for each sample basin (Fig. 2b). Mean precipitation rates observed between 1971 and 2000 were calculated for each basin using an 800 m resolution time series of precipitation rates (http://www.prismclimate.org).

We calculated potential Quaternary denudation rates from valley widening since the onset of pervasive glacial conditions since Quaternary. To estimate the impact that valley widening may have on Quaternary denudation rates, we calculated the rock mass that was removed from glacial processes. Two end-member models that may be used to compute the mass removed from the valleys due to widening by glacial processes are shown in Supplementary Fig. S6. The volume removed from glacial processes depends on the area of valley bottoms and relief. Since the height of Northern Cascades seems to be consistent since 15 Ma, we assume that fluvial and glacial valley relief are roughly equal. To calculate the volume of the rock removed by valley widening, we used the fraction of area in valley bottom and averaged local relief with a 10 km diameter window along points within the valley bottoms. Depending on the geometry of glaciation...
valley cross section, the volume can be varied (Supplementary Fig. S6). The lower end-member volume removal will be one half multiplied by mean areal fraction and mean relief, and the higher end-member will include mass removal in the side wall and ridge top. Mass from parallel retreating of side wall will be multiplication of mean areal fraction and mean relief, and ridge top will be one fourth of multiplication of mean areal fraction and valley width (assumed width = ~ 1 km on average). This removed volume is divided by 2.6 Ma (the minimum time at which glacial erosion was likely present in the Cascades), which will give the denudation rates from valley widening. Valley widening has plausibly enhanced denudation rates across the range during the Quaternary, which show a range of 0.005-0.047 mm/yr. Adding the denudation rates from valley widening and long-term denudation rates give the increased Quaternary denudation rates, but these rates are much less than our measured thousand-year timescale rates.

Third, we examined the potential biases of our measured denudation rates from previous glaciations. The glacial abrasion and ice cover could make the surface concentration of $^{10}$Be zero, which will result in an increase in apparent denudation rates. We used the analytical solution for concentration of cosmogenic nuclide with a function of time and depth to calculate the transient response of denudation rates after deglaciation$^5,19$. Two samples with highest (WR11) and lowest denudation rates (WR8) were selected for examples (Supplementary Fig. S4). The apparent denudation of WR11 falls within the error range of steady state denudation rates after 3 ka, while the apparent denudation of WR8 reaches the steady state rate approximately 17 ka after deglaciation. Because our study area was deglaciated 11-17 ka ago, the potential bias of denudation rate is 11-28 % in WR8 and 0 % in WR11. Therefore, samples with low denudation rates that might be affected by previous glaciations should be treated as ones that record maximum rates. However, the bias does not substantively change the general trend of denudation with other factors throughout these basins.

Denudation rates may also appear artificially higher due to the input of low $^{10}$Be concentration sediments from glacially reworked deposits or local, individual bedrock landslides. In this case, variations in $^{10}$Be concentration do not represent the variations in basin-average denudation rate. Instead, they will represent the degree of dilution resulting from the input of those sediments. A mass balance calculation shows that average of ~63%, and maximum of 87% (WR11), of the sediments would need to be sourced from these low $^{10}$Be concentration areas to explain the observed variation in erosion rates. We examined the possibility of sediment influxes from reworked glacial deposits. Previous studies in the North Cascades in Canada suggest that most of the sediments are derived from reworked glacial deposits$^20$. This conclusion is supported by the positive correlation between drainage area and sediment yield. However, in our study area, we could not observe such correlation. There are no correlations between predicted amount of diluting sediments and either drainage area or % area of post-glacial deposit$^2$ (Supplementary Fig. S5; Table S2). In fact, only 5 sample sites have glacial deposits whose area is larger than 1 % of total area. For other samples, though the sediments are predicted to be diluted for 35-82%, there are no observable sources for these diluting sediments. Furthermore, if the low concentration sediments indeed come from the postglacial deposits, a mass balance suggests that the denudation rates in those post-glacial deposits should be at least 40-200 times higher than hillslope erosion rates. Assuming this condition has persisted since ~ 10 ka, the amount of erosion corresponds to the lowering of 40-200 m of glacial deposits,
which is unrealistic considering the tens of meter thickness of postglacial deposits in Skykomish river\textsuperscript{21}. Moreover, a previous sediment modeling study in Rainy Creek, which is close to WR5, suggests that the main sources of sediment generation are mass-wasting processes\textsuperscript{22}. The observed landslide rate from repeat aerial photographic mapping is \( \sim 0.19 \) mm/yr\textsuperscript{22}, which is close to 0.28 mm/yr of WR5. In addition, studies in other deglaciated landscapes also suggested the mass wasting processes as significant sources of sediments\textsuperscript{23,24}.

We also consider the characteristic features in deglaciated topography in calculations of denudation rates and basin averaged parameters. Deglaciated landscapes have features that are different from fluvial landscapes, especially with regard to the fact that the sources of the sediment are not uniformly distributed throughout the basins. Therefore, we need to exclude sub-basin areas that do not contribute to the sediments of our samples. To this end, two scenarios were explored. First, we excluded watersheds that drain into glacial lakes, and assume that sediment in the size fraction analyzed (250 – 710 \( \mu m \)) is impounded in these features. Using land cover data from USGS\textsuperscript{25} and Google Earth imagery, we identified 167 lakes beyond which further sediment impoundment in lakes was not observed. Supplementary Fig. S1 shows the location of the lake outlets and their upstream drainage areas. The average fraction of excluded drainage area is \( \sim 9\% \), with individual basins containing between 0\% (WR2) and 44\% (MH17) of pixels subject to sediment impoundment in lakes. The denudation rates recalculated based on scaling factors that result from these adjustments show 0.08-0.55 mm/yr, which differs by 0\% to -18\%. Changes in denudation rates are less than 10\% in most of samples except for MH17 (-18\%). The recalculated parameters (precipitation, local relief with 5 km diameter, elevation, slope, channel steepness) differs < 10\%. This correction for glacial lake impoundment of sediment does not significantly alter the relationship between denudation, and the topographic and climatic parameters investigated.

Second, we explored another scenario, in which we excluded both the pixels in the flat bottom of glacial valley in addition to drainage areas of channels feeding glacial lakes. The rationale of this scenario is that low-gradient valley bottoms will not constitute a significant sediment source (although bypass will likely result along the low-gradient channel reaches). Thus, including the low gradient valley bottoms would produce misrepresentative parameter values both for our calculation of production rates and basin characteristics. To exclude the glacial valley bottom, we selected the pixels with slopes less than 5\(^\circ\) using the DEM\textsuperscript{17}. We exclude only the individual pixels with slope less than 5\(^\circ\), not all the upstream pixels because the low gradient reaches of these channels likely still carry the 250-710 \( \mu m \) sediment as suspended load. The area with slope less than 5\(^\circ\) is denoted in Supplementary Fig. S1a. This analysis results in exclusion of 4-46\% of area, an additional exclusion of 2-7 \% of the area after glacial lake correction. The denudation rates changed from 4 to -16 \% relative to uncorrected rates, which are slightly larger than rates of lake correction because the flat-bottomed areas of the landscape with low production rate are excluded. The climatic (precipitation) and topographic (relief, slope, elevation) parameters changed less than 10\%. The most significant change occurred in the calculation of channel steepness, since some of the channel points are located in the flat bottom of valley. The exclusion of points with slope less than 5\(^\circ\) increased the average channel steepness by 22-94\% in $Ksn$ with area > 3 km\(^2\). For comparison, mapped location of $Ksn$ with area > 3 km\(^2\) is shown in Fig. 1e and $Ksn$ with area > 3 km\(^2\) and
slopes $>5\,^\circ$ in Supplementary Fig. S1c. The calculated basin-averaged channel steepnesses with two corrections are shown in Supplementary Fig. S2. While the magnitude of $K_{sn}$ appears to be sensitive to this change, the relative partitioning of $K_{sn}$ between basins (and its relationship to climatic parameters) appears robust, despite the masking that we now use.

Fourth, we calculated two topographic metrics to examine the role of mass transport mechanisms that may be responsible for the observed enhancement in denudation rates. First, we calculated catchment averaged channel steepness index which likely increases with the rate of river incision. The longitudinal profile of channels can be described by a power law relationship between local channel slope ($S$) and drainage area ($A$):

$$S = K_s A^{-\theta}$$

(2a)

Where $K_s$ and $\theta$ are referred to as the channel steepness and concavity, respectively$^{27,28}$. Inspection of equation (2a) indicates that when $A$ and $S$ are plotted against one another in logarithmic space, $K_s$ and $\theta$ can be extracted by linear regression of these topographic attributes$^{27}$. The measurable topographic attributes, $K_s$, can be related to spatial distribution of denudation ($\varepsilon$): $K_s = \left(\varepsilon / K\right)^{1/n}$

(2b)

where $m$ and $n$ are empirical constants whose ratio ($m/n$) is close to 0.5$^{28,29}$ and $K$ is an erodibility factor that subsumes, for instance, the effects of rock erodibility and precipitation$^{28}$. In addition, fixing $\theta = \theta_{ref}$ and assuming constant $n$, we could calculate normalized channel steepness index ($K_{sn}$) which can be used to understand the relationship between topography, denudation, and erodibility factor. In a qualitative sense, $K_s$ increases with increasing channel slope at a reference drainage area, and so can be used to understand spatial changes in channel slope that cannot be explained by the changes expected as basin area changes.

We extracted $K_s$ for basins of the Washington Cascades using the 30 m National Elevation Dataset Digital Elevation Model (DEM). The DEM was processed to remove artifacts and impose continuous flow using the RiverTools software package$^{30}$. For this corrected DEM, we calculated upstream area and flow paths. We then used the raw DEM to extract channel profile segments whose points shared a common Strahler order. For each segment, local channel slope was calculated by determining elevations at an equal vertical interval (20 m) and applying a 2nd order finite difference approximation of the channel slope at these points$^{31}$. Next, flow area was extracted for each discrete point at which channel slopes were determined. This process was recursively repeated for every link of the channel network. In order to calculate the $K_s$ which representing only river incision, we also calculated $K_s$ using the channel networks with an area $> 3$ km$^2$, which effectively screened low order tributaries from our analysis to avoid debris-flow-dominated areas of the landscape$^{32}$. Finally, $K_s$ for each point was calculated using $0.4 < \theta < 0.6$, representing the likely range of $m/n$ due to fluvial bedrock incision process, which is the potential the rate-limiting process that sets denudation rates in this landscape since deglaciation$^{29}$. As shown in the text, this is the timescale represented by our cosmogenic denudation rate measurements (1.1-7.4 kyr). We show $K_s$ values for $\theta = 0.5$ with area $> 3$ km$^2$ and slope $> 5^\circ$ in Fig. 3a for illustration and present $K_s$ with area $> 3$ km$^2$ for $\theta = 0.4$, 0.5 and $\theta = 0.6$ in Supplementary Fig. S2 with two correction results,
exclusion of lake drainages and valley bottoms with slope $< 5^\circ$ vs lake drainages — all concavities explored share similar trends. Catchment averaged steepnesses were calculated by averaging steepness measurements. In recently glaciated landscapes, the channel steepness would not necessarily scale with spatial distribution of rock uplift rates because rapid glacial erosion processes denude the landscape at a rate that is largely insensitive to slope\textsuperscript{17,33} However, in fluvially dissected terrains, channel steepness is correlated with denudation from river incision\textsuperscript{34} (Supplementary Fig. S3). In the Cascades, the channel steepness shows high values across the range regardless of the variation of denudation and also shows a poor correlation with denudation rates, which suggests that enhanced denudation in this deglaciated landscape is not derived from increased river incision.

In order to examine the role of shallow landslides on denudation rates, we calculated the relative failure index which is a version of the infinite slope approximation for failure of a cohesionless frictional material coupled to a steady–state hydrologic model\textsuperscript{35}. The failure index is similar to the inverse of a-factor-of-safety values. For every flow path with an area $< l$ km$^2$, we calculated the wetness $W$, which is the ratio between hydraulic flux at a given rainfall amount relative to the degree of soil saturation:

$$W = \frac{qA}{bT\sin \theta}$$

where $W$ is wetness whose value is limited between 0 (unsaturated) to 1 (completely saturated), $q$ is the steady-state precipitation during a specific storm event, $A$ is contributing area draining across $b$ the contour length, $T$ is soil transmissivity when saturated, and $\theta$ is the local slope. Based on this wetness value, we calculate the slope failure index ($F$),

$$F = \tan \theta \left(1 - W \frac{\rho_w}{\rho_s}\right)^{-1}$$

(4)

where $F$ is failure index, $\phi$ is the internal friction angle, $\theta$ is the local slope, and $\rho_s$ is a wet bulk density of soil and $\rho_w$ is bulk density of water. Slopes were calculated using a second-order finite difference approximation of the slope between points extracted at equal vertical spacing along the down-slope flow paths in the DEM. This was to reduce the impact of local aberrations in that result from artifacts in the DEM when calculating averaged slopes. When $F$ is larger than 1, the combined effects of the local topography and hydrology are predicted to create failure of the shallow subsurface materials. This failure model of a cohesionless frictional material is applicable to our study area considering the small extent of barren land in the slopes between the saturated and unsaturated failure angle: ~ 7% of those slopes are in the barren land, while others are in soil-mantled landscapes\textsuperscript{25}. Assuming spatially uniform $T$ ($1 \times 10^{-4}$ m$^2$/s), $\rho_s$ (2.0 g/cm$^3$), $\rho_w$ (1.0 g/cm$^3$), tan $\phi$ (~1), and $q$ (mean annual precipitation), we calculate the failure index for each point in our catchments. We use the mean annual precipitation for net rainfall rate $q$. Since the monthly precipitation closely mimics the spatial pattern of annual precipitation ($R^2 > 0.98$)\textsuperscript{14}, we assume that the relative value of the “representative” steady precipitation used in the steady-state hydrologic model described above is represented by the spatial distribution of the mean annual precipitation. In this case, the magnitude of higher frequency events that likely trigger landsliding scales (in
space) with the mean annual precipitation, and as such, can be mapped into the effective (and unknown) value of T. The pointwise failure indexes are calculated for channel points with area < 1 km², and the fraction of points expected to be dominated by landslides (failure index larger than 1) was counted for each basin (hereafter referred to as the failure fraction. Denudation rates increase systematically with the failure fraction (Fig. 3b), which suggests that the instability on the slope by shallow landslide is a plausible mechanism for the enhanced denudation rate. In order to account for uncertainty in transmissivity and the relationship between the mean annual and steady-state precipitation during a failure episode, we varied q/T over two orders of magnitude and recomputed the relative failure index for each of the basins. Basin averaged failure index and failure fractions with variation of T are shown in Supplementary Fig. S7. In addition, to evaluate the relative contributions of slopes and precipitation to slope failure, we performed sensitive analyses for failure index calculation. We, first, calculated failure index with varying precipitation across the range with slope fixed to the mean value (0.52), and then varying slope across the range while precipitation rate was fixed to its mean value (1780 mm/yr). The analyses show that both are likely important in triggering slope failure, but denudation rates correlate most strongly with the failure fraction when is imposed the precipitation gradient with slope fixed, and secondarily to the failure fraction when is imposed slope variation with precipitation fixed (Supplementary Fig. S7). This suggests that increases in pore pressure due to current rainfall gradients across the Cascades, and secondarily the pre-glacial variations in slopes across the range cause the shallow landslides to denude the basins in the west more vigorously than their dryer counterparts. We also calculated the critical rainfall rates (qcr) which represent the minimum steady state rainfall which induces the shallow landslides. The critical rainfall rates represent a sole effect of topography on slope failures for those points in conditional slopes between 30° to 45°. The basin-averaged critical rainfall rates show a weak inverse relation with denudation (R² = 0.330; Supplementary Fig. S8), which also suggests that more frequent failure due to lower critical rainfall thresholds might increase denudation rates. In addition, a previous study in our study area and other studies in deglaciated area also suggest that mass-wasting processes (shallow and deep landslides) could generate higher denudation rates than hillslope erosion 22-24.

Finally, we examined whether the rates we measured that quantify denudation over thousand-year time-scales capture a tectonic acceleration of rock uplift rates, rather than recording the effect of recent deglaciation. In particular, our ¹⁰Be data show the denudation rate during the past ~1.1-7.4 ka is 4 times higher than the Myr-timescale denudation rates derived from thermochronology. It is possible that these high denudation rates may have persisted for longer periods of time and predate the post-glacial recession. Given that there are few methods that can resolve the denudation rate history in the interval between the apatite (U-Th)/He cooling ages and the cosmogenic derived denudation rates, our study cannot rule out the possibility that tectonic rock uplift rates have recently accelerated here. However, there are several independent lines of evidence that suggest that is not likely the case: 1) if these rapid denudation rates span prior to ~2.6 Ma (1.5 km/0.57 mm/yr), the (U-Th)/He cooling ages would have had younger age than 2.6 Ma; but this is not observed. 2) According to a previous study 18, the high relief topography of the northern Cascades, including our study area, seems to be consistent since Miocene (~15 Myr). If large changes in tectonic rock uplift rates
occurred over the last several million years, it is unclear why the topography would not have become higher as topography likely would have had ample time to adjust to this long-term change. 3) There is limited evidence for recent tectonic activity in the Cascades—active Quaternary faults and seismic activity in this study area are sparse, and we could identify no candidate structures that would facilitate such rapid tectonic rock-uplift in the brittle upper crust of the Cascades (http://earthquake.usgs.gov/hazards/qfaults), although exposure of geologic structures in this landscape may obscure identification of these features. 4) In contrast, the topography of the Cascades can be identified as remnants from glacial erosion, indicating that the landscape form is currently imbalanced with respect to the processes that now erode it\textsuperscript{17,36}. The steep side valley walls in glacial landscapes tend to be steeper than their fluvial counterparts, and mass wasting accelerates with precipitation and landscape steepness. It would be unusual if such an oversteepened landscape did not have a higher denudation rate than might be expected in a fluvially adjusted landscape\textsuperscript{23,24,37}. In fact, several studies pointed out that climatically perturbed landscapes show enhanced short term denudation rates relative to long-term averages, which suggests the increased short term variations in denudation rates are due to a transient response to glacial and interglacial cycles\textsuperscript{23,37-40}. Therefore, while there are no chronologic methods available to rule out the possibility that rock uplift rates have increased recently, the consistency in relief of the range over geologic time, the lack of candidate structures to accommodate the posited modern rapid tectonic uplift, and the rapid erosion of the steep side-valley slopes left behind by glaciers provide independent, albeit incomplete, lines of evidence that argue against an acceleration of tectonically-driven rock uplift rates over the Plio-Quaternary.
Supplementary Table S1. Summary of basin characteristics and denudation rates.

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Latitude (°)</th>
<th>Longitude (°)</th>
<th>Drainage area (km²)</th>
<th>Correction for lake lake + slope areal fraction</th>
<th>Precipitation (mm/yr)</th>
<th>Long-term denudation (mm/yr)</th>
<th>Slope (grad.)</th>
<th>Local relief (&gt; 3 km)</th>
<th>Mean elevation (m)</th>
<th>Mean log10(Ksn)</th>
<th>Failure index (log10(FI))</th>
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<tbody>
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<td>WR1</td>
<td>47.821</td>
<td>-120.422</td>
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<th>Sample Name</th>
<th>Production rate (atoms/g/yr)</th>
<th>Scaling Factor</th>
<th>Grain size (μm)</th>
<th>Mass of Quartz (g)</th>
<th>Amount of Be carrier (mg)</th>
<th>Spallation linearity</th>
<th>Muons shielding</th>
<th>Amount of 10Be/9Be (×10^13)</th>
<th>10Be/9Be concentration (×10^13 atom/g)</th>
<th>Denudation rate (mm/yr)</th>
<th>Apparent age (yr)</th>
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<td>0.3297</td>
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a calculated from 30 m DEM
b long-term Myr-timescale denudation rates from apatite (U-Th)/He ages\textsuperscript{10,11}
c mean local relief calculated following window size
d channel steepness with reference concavity 0.5
e failure index calculated with $T = 10^{-4}$ m$^2$/s
f constant (time-invariant) local production rate based on Lal/Stone scheme\textsuperscript{5,6}. Shielding factors are not applied.
g correction for topographic non-linearity including latitude-altitude scaling factors and pressure correction\textsuperscript{6}
h topographic shielding\textsuperscript{8}
i $^{10}$Be/$^9$Be ratios were measured using calibration method\textsuperscript{2} and corrected for blank combined with analytical and blank error ($^{10}$Be/$^9$Be = 0.63 ± 0.16 × 10$^{-14}$)
j Denudation rates are calculated from CRONUS-Earth online calculator version 2.2\textsuperscript{7}, time-dependent corrections of Lal/Stone scheme\textsuperscript{5,6} assuming rock density 2.6 g cm$^{-3}$
k the time spent in the uppermost 60 cm of rock surface.

Supplementary Table S2. Areal fraction for categories of land cover, lithology, and landslides.

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Land Cover$^a$</th>
<th>Postglacial Deposit$^b$</th>
<th>Maximum Glacier Extent$^c$</th>
<th>Landslides$^d$</th>
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</thead>
<tbody>
<tr>
<td>WR1</td>
<td>Water</td>
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<td>0.00</td>
<td>0.07</td>
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<tr>
<td>WR3</td>
<td>Vegetated; Natural</td>
<td>0.00</td>
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<td>0.04</td>
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<tr>
<td>WR4</td>
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<td>0.00</td>
<td>0.09</td>
</tr>
<tr>
<td>WR5</td>
<td>Shrubland</td>
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\textsuperscript{a} land cover dataset\textsuperscript{25}. The fractions will be summed to 1
\textsuperscript{b} postglacial deposits including Pleistocene glacier drift, till, outwash, and moraine\textsuperscript{2}
\textsuperscript{c} percentage of area within maximum glacier extent\textsuperscript{17}
\textsuperscript{d} percentage of area with landslides\textsuperscript{41}
Supplementary Fig. S1. a) Map of areas used for denudation rate correction. Red region indicates glacial lake drainages with yellow points of glacial lake location, and green region indicates areas with slope less than 5°. b) local relief calculated from a 10 km-diameter-window. c) channel steepness ($K_{sn}$) with area > 3 km² and slope > 5°.

Supplementary Fig. S2. Plot of channel steepnesses and denudation rate (mm/yr). Channel steepnesses based on reference concavity 0.4, 0.5, and 0.6 from channel networks with an area > 3 km². Channel steepnesses from two corrections are shown for comparison (a-c) channel steepness corrected both glacial lake drainages and valley bottoms (slope < 5°) (d-f) channel steepness corrected only for glacial lake drainages. Error bars represent 1 $\sigma$ standard deviation in $K_{sn}$ within sample basins, and 1 $\sigma$ uncertainty in measured $^{10}$Be concentration in denudation rates. Uncertainties in regressions denote 1 $\sigma$ residual error ($\sigma_r$) of linear fit.
Supplementary Fig. S3. Plot of $^{10}$Be denudation rate and catchment average channel steepness data from this study and San Gabriel Mountain (SGM)\textsuperscript{34}. For comparison to SGM, our $K_{sn}$ is recalculated with reference concavity = 0.45 and averaged over points with area > 3 km\textsuperscript{2}. Data represented by blue circles are from fluvial catchments in the San Gabriel Mountains, and points in square come from the Washington Cascades (grey only corrected for glacial lakes, and black corrected for both glacial lakes and valley bottom (slope < 5\degree). $K_{sn}$ values for the Cascades are high even for the low denudation rates, suggesting that oversteepened slopes from previous glacial processes are still imprinted in the landscapes. The dark blue line shows the non-linearity of the denudation rate and $K_{sn}$ values, resulting from the stochastic threshold model for SGM\textsuperscript{34}.

Supplementary Fig. S4. Changes of apparent denudation rate since deglaciation. The apparent denudation rate since surface zeroing by glacier abrasion and ice cover was denoted by solid line, and the steady state denudation rates by dotted line. Two exemplary samples were chosen (WR11 with highest denudation and WR8 with lowest denudation). Considering the deglaciation started around 11-17 ka, the potential bias of denudation rates is not significant enough to change the trend of dataset.
Supplementary Fig. S5. plot of area (km$^2$) and % area of postglacial deposit versus % sediments from the sources with low $^{10}$Be concentration (diluting sediments) (see Supplementary Table S2).

Supplementary Fig. S6. The simple geometry used for removed mass volume from valley widening ($W_v$: valley width and $R$: local relief (10 km)). The black line shows the glacial valley cross-section and blue line for fluvial valley cross-section. The light blue region is the removed cross-sectional area (a) for lower end-member mass removal and (b) high end-member mass removal including parallel retreat of side wall and reduction of ridge top.
a) Real distribution

b) Precipitation gradient

c) Slope gradient
Supplementary Fig. S7. Denudation with basin averaged failure index and failure fraction with varying $T = 10^{-2}, 10^{-3}, 10^{-4}, 10^{-5}$ m$^2$/s. For comparison, calculated failure index and failure fraction with varying precipitation across the range with slope fixed to mean value are shown in (b), and those with varying slope across the range with precipitation fixed to mean value is shown in (c).

Supplementary Fig. S8. Plot of relative critical rainfall rate versus denudation (mm/yr) with $T = 10^{-4}$ m$^2$/s. Smaller value in critical rainfall indicates that it requires less rainfall to induce failure, which appears to be related to the increase of denudation rates.

41 Washington Division of Geology and Earth Resources. Landslide Hazard Zonation work products (landslide24k) (Washington Division of Geology and Earth Resources Olympia, Washington 2008).