Himalayan valley-floor widths controlled by tectonics rather than water discharge

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Article

Keywords: valley widening, channel steepness, tectonics, exhumation

Posted Date: October 11th, 2022

DOI: https://doi.org/10.21203/rs.3.rs-2065309/v1

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Himalayan valley-floor widths controlled by tectonics rather than water discharge

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Abstract

Himalayan rivers transport $\approx 10^3$ Mt of sediment annually to ocean basins. River valleys are an important component of this routing system: while sediment is stored in valleys, signals of climate change and erosional patterns can be modified or even destroyed. Despite a critical need to understand the spatial distribution, volume and longevity of these valley fills, controls on valley location and geometry are unknown, and estimates of sediment volumes are based on never-tested assumptions of valley widening processes. Here we extract 1,644,215 valley-floor width measurements across the Himalaya to determine the dominant controls on valley-floor morphology for the first time, and to test underlying assumptions of sediment storage volumes. We use
random forest regression to estimate the importance of potential con-

trolling variables, and find that channel steepness, a proxy for rock

uplift, is a first-order control on valley-floor width. We also analyse

a novel dataset of 1,797 exhumation rates and find that valley-floor

width decreases as exhumation rate increases. We therefore suggest that

valley-floor width is adjusted to long-term tectonic exhumation rather

than being controlled by water discharge or bedrock erodibility, and

that valley widening predominantly results from sediment deposition

along low-gradient valley floors rather than lateral bedrock erosion.

Keywords: valley widening, channel steepness, tectonics, exhumation

Valleys in mountain systems act as transient depositional sinks for sediment on

its journey from its source on mountain hillslopes to its final resting place in the

foreland or ocean basin. This storage within mountain systems has important

implications for our ability to reconstruct external signals from sedimentary

archives, with these signals potentially being buffered, shredded, or destroyed

en route [1–3]. Therefore, understanding the spatial distribution, volumes, and

longevity of valley sediment fills is essential if we want to accurately reconstruct

mountain landscape evolution. However, controls on the spatial distribution of

valley fills across the Himalaya are currently unknown. Past efforts to map the

volumes and residence times of valley fills across the orogen at scale [4] rely

on the assumption that topography underneath the valley surface is similar to

that of the exposed valley side-slopes, and therefore that little lateral erosion

of the valley walls has taken place. This assumption has never been tested on

an orogenic scale.

We consider two end-member conceptual models of how valleys may widen

in mountain systems (Figure 1). One end-member, the ‘bedrock’ model,

assumes that valley-floor width changes occur through lateral erosion of the

valley walls and the balance between vertical incision and lateral erosion.

Bedrock wall erosion is likely to occur when the channel is frequently in con-
tact with the bedrock walls [5, 6], such as in narrow valleys with little-to-no

sediment deposition. This model of lateral erosion is similar to the detachment-

limited model for vertical incision, which is commonly used in mountain

landscapes [e.g. 7]. Valley-floor width $W_v$ [L] in the bedrock model has been

suggested to scale with bankfull water discharge $Q_w$ [L$^3$ T$^{-1}$], modulated by

an erodibility coefficient $K_v$ that reflects the impact of lithology [e.g. 8–14]:

\[ W_v = K_v Q_w^{c_v}. \] (1)

In landscapes where recent changes in rock-uplift rate cause transient

adjustment of fluvial systems, this relationship has been shown to break down

[e.g. 15, 16]. An alternative formulation in tectonically active regimes suggests

that valley width is also dependent on valley slope ($S$) [6, 15] (Supplementary

Information).
Despite its fairly common application, the bedrock model is contradicted by field observations, which show that mountain valleys tend to be infilled with sediment (Figure 1). In valleys with significant sediment deposition, widening through lateral erosion of the bedrock walls is only likely to occur if lateral erosion rates greatly exceed vertical erosion rates, such that the channel regularly moves across the valley floor, impinging upon the sidewalls [5, 6]. The alternative ‘alluvial’ model assumes that width changes in sediment-filled valleys occur primarily through sediment deposition and/or erosion. If we imagine a roughly V-shaped valley infilled with sediment (Figure 1), then increasing sediment fill would widen the valley, whereas incision into the fill would narrow it. $W_v$ would be set by the depth of sediment in the valley and the angle of the surrounding hillslopes.

These end-member models represent contrasting mechanisms of valley-floor width changes, which are likely to be controlled by different factors (Figure 1). In both cases, rock uplift is likely to be an important control on $W_v$, because high uplift rates elevate channel slopes, increasing flow velocity and resulting in narrowing [15]. Alternatively, increased frequency of landsliding in regions of high uplift [e.g. 17] could block river channels, inducing upstream alluviation and valley widening. However, $Q_w$ and $K_v$ are likely to be important controls on valley-floor width [8, 11, 13] only in the bedrock model. In a valley that changes width only due to sediment erosion or deposition, variations in $K_v$ linked to lithology are unlikely to play a dominant role, as the bedrock walls are never eroded. Faulting may increase rock fracturing and therefore erodibility [e.g. 18, 19]: we might therefore expect that valleys in highly fractured zones (such as near seismogenic faults) would be wider in the bedrock model, but not in the alluvial model. The ratio of sediment flux to water discharge, $Q_s/Q_w$, rather than $Q_w$ alone, is likely to influence valley-floor width in the alluvial case. Field studies [20, 21] and physical experiments [22] have demonstrated that a decrease in the ratio of $Q_s/Q_w$ leads to incision, terrace formation and valley narrowing, whereas an increase in the ratio of $Q_s/Q_w$ leads to sediment deposition and valley widening. Over orogenic scales, we can therefore hypothesise that a positive correlation between $W_v$ and $Q_w$ would not exist if widening primarily occurs due to sedimentation.

Sediment storage volumes estimates across the Himalaya [4] implicitly use the alluvial model, because they assume that little bedrock erosion of the valley walls is occurring to modify the topography of the valley floor. In this contribution, we attempt to distinguish dominant controls on $W_v$ across the Himalaya and test these end-member models of valley widening and sediment storage volumes on an orogenic scale. We generate the first dataset of valley-floor widths across the Himalaya and investigate the relative importance of key hypothesized controls on valley-floor width through random forest regression. We also explore links between $W_v$, $k_{sn}$, and exhumation rate using a novel dataset of thermochronometric cooling ages from across the Himalaya [23].

We extract $W_v$ from every major river basin in the Himalayan orogen using a new method for automatically extracting widths of identified valley floors.
Himalayan valley-floor width

Fig. 1 End-member models of valley widening mechanisms and different factors that may control widening in each model. The photographs show examples of the two end-member valley types in the Kedarnath Valley, Upper Ganga basin (photo credit R. Devrani).

<table>
<thead>
<tr>
<th>Controlling factor</th>
<th>Bedrock valleys</th>
<th>Alluvial valleys</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rock uplift</td>
<td>High uplift leads to increased vertical erosion compared to lateral and narrowing</td>
<td>High uplift causes increased channel slopes. This increases transport capacity, causes sediment erosion and narrowing</td>
</tr>
<tr>
<td>Erodibility</td>
<td>High erodibility leads to enhanced lateral erosion and wide valleys</td>
<td>Erodibility does not impact valley width</td>
</tr>
<tr>
<td>Active faults</td>
<td>Increased fracturing in seismogenic zones may cause enhanced lateral erosion and widening</td>
<td>Bedrock fracturing does not impact valley width. Enhanced sediment supply in seismogenic zones may increase deposition and cause widening</td>
</tr>
<tr>
<td>Water discharge ($Q_w$)</td>
<td>Higher $Q_w$ leads to enhanced erosion of walls and widening</td>
<td>Dependent on ratio of $Q_s$ to $Q_w$ (see below)</td>
</tr>
<tr>
<td>Sediment discharge ($Q_s$)</td>
<td>Intermediate $Q_s$ enhances abrasion leading to widening (tools and cover effect)</td>
<td>High $Q_s/Q_w$: sediment deposition and valley widening. Low $Q_s/Q_w$: erosion and valley narrowing</td>
</tr>
</tbody>
</table>

[24, 25]. We filter the dataset to remove intersecting width measurements and modern glaciers (Methods): the resulting dataset contains a total of 1,644,215 width measurements. We then grid $W_v$ into 10 km pixels to better reveal spatial trends: Figure 2 shows the distribution of $W_v$ across the orogen. We
quantify each of the controlling factors that may affect $W_v$ outlined in Figure 1 (Methods).

![Fig. 2](image)

**Fig. 2** (a) Map of the Himalayan orogen showing basins used for width analysis; (b) Topography across the region with main structural boundaries: MFT = Main Frontal Thrust, MBT = Main Boundary Thrust, MCT = Main Central Thrust, STD = South Tibetan Detachment; (c) distribution of valley-floor width; and (d) distribution of normalised channel steepness ($k_{sn}$) across the Himalaya. The data in (c) and (d) are gridded into cells with 10 km spatial resolution.
**Controls on valley-floor width**

Figure 3a shows that elevation across strike is an important factor in determining valley-floor width. We find a bimodal distribution of valley-floor width with elevation, where valleys are widest at elevations <1000 m and >4000 m. We would expect the southern, external parts of the orogen at lower elevations to have wider valleys as discharge increases toward the southern foreland. Although we remove modern glaciers from our analysis, widening at high elevations is likely a result of Quaternary glaciation causing extensive erosion: glacially-scoured valleys are typically wider than fluvially carved valleys [26].

Figure 3b also shows that there is variation in median valley-floor width among the main tectono-stratigraphic units. This variation is possibly due to lithological control on valley widening, as the narrowest valleys are found in the high-grade metamorphic gneisses and granites of the Greater Himalayan Sequence (GHS) [27]. The widest valleys are found in the unmetamorphosed sediments of the Tethyan Himalayan Sequence (THS), sourced from the passive margin of the Indian continent; and the sedimentary units of the Siwalik foothills in the Sub-Himalayan Zone (SHZ). However, these variations with tectono-stratigraphy coincide with the trends with elevation discussed above, making it difficult to separate these two factors. Figure 3e shows that the widest valleys occur at the farthest distances from the major tectonic structures (MFT, MBT, MCT, or STD). This suggests that fracturing is unlikely to be controlling valley-floor width, as we would expect to find the widest valleys closest to major faults if increased rock erodibility through fracturing [19] were resulting in enhanced valley wall erosion.

Rock-uplift rates across the Himalaya since the middle Miocene have been controlled primarily by the geometry of the Main Himalayan Thrust (MHT) [28], a northward-dipping décollement that separates the Indian and Eurasian plates and is the basal detachment for the MFT, MBT, and MCT. The MHT is suggested to be relatively flat under much of the LHS, steeper to the north over a mid-crustal ramp [e.g. 29] beneath the GHS, then flat again beneath the THS (Figure 4). The ramp is associated with faster rock uplift rates and steeper topography [30], with a ‘physiographic transition’ marking the change from the southern (shallower) flat to the ramp. Our results show a distinct area of wide valley floors within the LHS in central Nepal (Figure 2b), with a sharp transition to narrow valleys north of the physiographic transition (Figure 4). Considering that the physiographic transition cuts across the LHS in this region, the flat-ramp-flat structure of the MHT appears to influence valley-floor width in central Nepal more strongly than the transitions across tectonostratigraphic units.

Models of valley widening through bedrock lateral erosion predict a monotonic relationship between water discharge and valley-floor width (equation 1). Our results do not show this relationship (Figure 3c). Although the widest valleys are found in regions with the highest water discharges, the narrowest valleys (86 ± 180 m) tend to coincide with intermediate discharges of 0.2 - 1.0 m$^3$ yr$^{-1}$. At the lowest water discharges of 0.01 - 0.05 m$^3$ yr$^{-1}$, the
median valley-floor width increases to 152 ± 192 m. This lack of correlation suggests that, in contrast to the commonly applied model of width evolution through lateral bedrock erosion, water discharge is not the dominant control on valley-floor width across the actively uplifting Himalayan region.

There is, however, a clear correlation between valley-floor width and $k_{sn}$ (Figure 3d). We tested this relationship by isolating for the influence of tectono-stratigraphic unit, and found that the correlation between $W_v$ and $k_{sn}$ is consistent across all individual units (Supplementary Information). To account for the competing influence of $Q_w$ and $S$ that may affect valley-floor width, we also calculated a discharge-weighted channel steepness, $k_{sn-q}$ [32]. We found that incorporating discharge into the calculation of $k_{sn}$ did not alter the observed relationships between channel steepness and valley-floor width (Supplementary Information). Normalised channel steepness is a widely accepted proxy for rock-uplift rate [e.g. 33, 34], suggesting that valley-floor width across the Himalaya responds to spatial variations in rock-uplift rate.

To further test this potential tectonic control of valley-floor width, we use a recent comprehensive dataset of 1,797 thermochronometric ages across the region [23] (Figure 5), from which we estimate exhumation rates using a simple 1D thermal model (Methods). Figures 5b and 5c show a correlation between valley-floor width, exhumation rate, and $k_{sn}$. The lowest exhumation rates of 0.1 - 0.2 mm yr$^{-1}$ correspond to the widest valleys and the lowest channel steepness indices. Intermediate exhumation rates between 0.3 - 0.9 mm yr$^{-1}$
show less variation in both valley-floor width and channel steepness respectively, whereas the highest exhumation rates $\geq 2$ mm yr$^{-1}$ have narrow valley floors and steep channels. Variations in exhumation rate in the Himalaya have been argued to also be strongly tectonically controlled [28, 35, 36]. Although several studies show that climate processes in the Himalaya are linked with exhumation [e.g. 42], we find no relationship between valley-floor width and mean annual precipitation (Supplementary Information). The positive correlation between $W_v$ and exhumation rate, along with the changes in $W_v$ across the flat-ramp-flat geometry of the MHT (Figure 4), suggests that valley-floor width across the Himalayas is likely controlled by tectonics.

**Importance of controls on valley-floor width**

Figures 3a to 3e demonstrate that there are many factors that vary with valley-floor width across the Himalayan orogen: we therefore take a data-driven approach to determine which has the strongest influence using random forest (RF) regression. RF regression is a form of supervised machine learning, which uses an ensemble of decision trees to predict a target variable (e.g., $W_v$) from a high-dimensional dataset [e.g. 37]. It allows the calculation of variable importance (VI): the most important variables used to predict the target variable. It does not require assumptions about the structure of the underlying
Fig. 5 (a) Map of exhumation rate derived from thermochronometry data across the Himalaya: the colours represent the exhumation rate in mm yr\(^{-1}\), symbols represent the thermochronometric system. (b) Boxplots showing relationship between valley-floor width and exhumation rate: the numbers above each box show the number of samples in the corresponding bin. (c) Boxplots showing the relationship between channel steepness \((k_{sn})\) and exhumation rate.  

Data, and therefore is useful in cases where the relationship between the target variable and the predictors is not known \textit{a-priori} [38]. To explore key controls on \(W_v\) we focus on the following variables based on the conceptual model shown in Figure 1: i) elevation; ii) \(k_{sn}\); iii) water discharge; iv) bedrock erodibility; and v) distance from the nearest fault (either the MFT, MBT, MCT or STD; Supplementary Information). We calculate erodibility using cosmogenic radionuclide (CRN)-derived erosion rates and \(k_{sn}\) (Methods).

We estimate VI in our RF regression using two different metrics (Methods), and found that \(k_{sn}\) is the most important predictor across all regression models (Figure 3f), with erodibility consistently the least important predictor. Elevation, water discharge and distance to the nearest fault have relatively similar importance, although elevation tends to be the more important variable out of these three. The relative importance of elevation potentially reflects the influence of glacially widened valleys at high elevations along with wide valleys at lower elevations, where channels begin to exit the mountain front.
Discussion and Conclusions

Our results suggest a moderate importance of \( Q_w \) and a low importance of bedrock erodibility on \( W_v \), contrasting with common models of valley widening by lateral erosion of bedrock walls (equation 1). We therefore suggest that valley-floor widening is more likely to occur via sediment accumulation in low-gradient valleys, corresponding to the alluvial model in Figure 1. This finding suggests little modification of the bedrock valley floor topography under these fills, and thus confirms a key assumption used to calculate stored sediment volumes across the orogen [4].

In contrast, we find that \( k_{sn} \) is a first-order control on valley-floor widths, and that both \( k_{sn} \) and \( W_v \) scale with exhumation rate. To our knowledge, this correlation has not been previously demonstrated. It suggests that the overall distribution of valley fills is driven by long-term tectonically controlled exhumation, rather than valley fills being randomly distributed, which we might expect if these deposits are formed by stochastic processes such as landsliding. Landslides have been suggested to occur more frequently in areas of rapid exhumation [17]: if landsliding was controlling valley-floor widths, we would expect more frequent landslide dams in rapidly exhuming regions and subsequent valley widening. This is not supported by our findings. Instead, we suggest that high rock-uplift rates in rapidly exhuming regions, reflected by high values of \( k_{sn} \), are likely to increase transport capacity, eroding sediment and therefore causing valley-floor narrowing.

This correlation also has important implications for the residence times of valley fill deposits: the apparent adjustment of \( W_v \) to the long-term exhumation rate measured over timescales of \( 10^5 - 10^6 \) years suggests either that individual valley fills persist over geological timescales, or that valley-floor width adjusts relatively rapidly to the local exhumation rate. Based on estimating sediment volumes, mean sediment aggradation rates, and valley trapping efficiencies, the residence time of Himalayan fills has been suggested to exceed \( 10^5 \) years for the largest valleys [4]. However, other studies have shown that flushing of valley fills and erosion of the valley floor can occur during extreme events [39–41]. If this is the case, our results suggest that re-filling of the valley to adjust to the local exhumation rate occurs on faster timescales than that of the next cycle of valley-fill removal.

The link between exhumation and \( W_v \) also has important implications for sediment routing systems and the transmission of seismic and climatic signals to sedimentary basins. If slower exhumation rates lead to wider valleys through the alluvial model, then sedimentary signals of external forcing in slowly exhuming areas are likely to spend more time in storage compared to rapidly exhuming areas, resulting in either buffering or shredding of the signal before it reaches its depositional sink [e.g. 2, 3]. Future work is needed to further explore i) the timescales of preservation of Himalayan valley fills; ii) the impact of exhumation rate on the propagation of allogenic signals; and iii) the sub-surface geometry of valley deposits to allow further investigation into valley widening mechanisms.
References


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Methods

Extraction of topographic metrics. Firstly, we isolated our analysis to the extent of the orogen [27, 43], including the tectono-stratigraphic units of the Sub-Himalayan Zone (SHZ), the Lesser Himalayan sequence (LHS), the Greater Himalayan sequence (GHS), and the Tethyan Himalayan sequence (THS) and excluding both the western and eastern syntaxial regions. We then split the DEM into major river catchments using catchment outlines from the Hindu Kush Himalayan region [45] and limited our analysis to those draining to the southern edge of the orogen. We then analysed valley-floor width for every major river basin, using a novel method for reproducibly and automatically extracting valley-floor width from digital elevation models (DEMs) [25]. This method first identifies floodplains using a threshold of slope and elevation above the nearest channel [24]. These thresholds can either be set manually by the user or defined automatically; to ensure consistency across the orogen we manually set a slope threshold of 0.15 and an elevation threshold of 100 m. The method then identifies the main flow direction of the channel and calculates valley-floor width orthogonal to this. The minimum possible width measurement is 60 m, which is set by the resolution of the DEM (2 DEM pixels). Following extraction of width measurements for every channel, we removed any measurements that intersected each other (i.e., at tributary junctions) from the dataset as these measurements are unlikely to represent the true valley-floor width. We also removed measurements from modern glaciers across the Himalayas using the glacier outlines from the Randolph Glacier Inventory (RGI) [48].

We calculated the elevation of each valley-floor grid cell using the Copernicus 30 m DEM, and determined the underlying tectono-stratigraphic unit using a geologic database [27]. We calculated normalised channel steepness ($k_{sn}$ (m^{-0.9})) across each river basin using a segmentation approach [49] as implemented in LSDTopoTools [50]. $k_{sn}$ is often used as a proxy for rock-uplift or erosion rates and has been shown to correlate with local relief and catchment-averaged erosion rate across the Himalaya [e.g. 34, 51, 52, 54–56]. We used a reference concavity value, $\theta = 0.45$, which has previously been estimated for the Himalayan region [32, 34, 58]. We gridded the $k_{sn}$ data using the same approach as for valley-floor width (Figure 2b).

To estimate water discharge, $Q_w$, we use a simple proxy based on weighting upstream drainage area ($A$) by mean annual precipitation ($P$) [32]:

$$Q_w = PA,$$
We estimated $P$ from 1981-2019 across the Himalaya using the Climate Hazards Group InfraRed Precipitation with Station (CHIRPS) dataset, which combines 0.05° resolution satellite imagery with ground-station data [59]. The advantage of using the CHIRPS dataset is that it has a near-global rainfall time series for more than 30 years, giving longer term estimates of $P$ that should be less sensitive to short-term temporal variations. We calculated $P$ from this dataset using Google Earth Engine, then resampled $P$ to a spatial resolution of 30 m to correspond to that of the topographic data. We test discharge rather than drainage area as the Himalaya have a strong orographic precipitation gradient resulting in an order-of-magnitude variation in $P$ across strike as well as an $\approx 6$-fold increase in rainfall from west to east [60, 61].

To investigate the potential impact of fracturing on bedrock erodibility we also calculated the Euclidean distance of each grid cell from the nearest major tectono-stratigraphic boundary (either the Main Frontal Thrust (MFT), Main Boundary Thrust (MBT), Main Central Thrust (MCT), or South Tibetan Detachment (STD)) [27].

Compilation of thermochronology data and calculation of exhumation rates. We updated an existing compilation of thermochronometric data from the Himalaya [62] to include more recent publications up to July 2022, including all data falling within the basins outlined in Fig. 2a. We include results from five thermochronometric systems in our analysis: apatite and zircon (U-Th)/He (AHe, ZHe) and fission-track (AFT, ZFT), and white-mica $^{40}$Ar/$^{39}$Ar (MAR). We removed any cooling ages $\geq 50$ Ma, as these ages are pre-Himalayan [43] and are therefore unrepresentative of valley-forming processes, as well as samples from the SHZ, as these are generally incompletely reset since deposition [63]. We also filtered the dataset based on uncertainty by removing any samples where the $1\sigma$ uncertainty in predicted exhumation rate was greater than the exhumation rate itself (Supplementary Information), and we removed any samples within the boundaries of modern glaciers as mapped by the RGI [48]. The complete dataset and associated references can be found in [23].

We use a 1D thermal model that assumes vertical exhumation and thermal steady state to estimate exhumation rates from the thermochronology data. The model (refer to [23] for details) takes into account the advective perturbation of the geotherm by rapid exhumation [65] and the control of cooling rate on closure temperature of each thermochronometric system [66, 67]. We use the sample elevation to estimate the surface temperature using a linear atmospheric lapse rate (5 °C/km) and a constant sea-level temperature (25 °C), as well as to estimate the vertical difference between the sample elevation and the average elevation smoothed within a radius that depends on the estimated closure depth of each thermochronometric system [68]. The latter is used to correct the estimated exhumation rate for relative sample elevation. For other model parameters, we assume the following: an initial linear geotherm of 25 °C/km, a thermal diffusivity of 30 km$^2$/Myr, and a model thickness of 30 km.
We then mapped each exhumation rate sample to the corresponding valley-floor width cell in the gridded 10 km dataset, and binned valley-floor width and $k_{sn}$ by exhumation rate.

**Erodibility index.** We calculated an erodibility index, $K$, for each of the main tectono-stratigraphic units across the Himalayan orogen using a compilation of catchment-averaged erosion rate data from cosmogenic radionuclides [69], similar to the approach of [70]. The commonly-used stream power incision model (SPIM) predicts a non-linear relationship between channel slope and erosion rates:

$$E = K A^m S^n,$$

which we can rearrange to find an expression for channel slope, $S$:

$$S = \frac{E^{1/n}}{K} A^{-\theta},$$

where $\theta = m/n$. We can simplify this equation to:

$$S = k_{sn} A^{-\theta},$$

$$k_{sn} = E/K^{1/n}. \tag{6}$$

We estimate $k_{sn}$ as described above, and then assume that the CRN-derived erosion rates are representative of erosion across the entire basin, such that for each point on the network, we know $k_{sn}$ and set $E$ as the catchment-averaged erosion rate. We can then rearrange equation 6 to solve for erodibility at each point on the channel network, $K_i$:

$$K_i = \frac{E}{k_{sn} n}. \tag{7}$$

Many studies have suggested through both numerical modelling and field studies that $n$ is likely to be $> 1$ [e.g. 70–72], with $n \approx 2$ suggested to be reasonable in most cases [73]. We therefore set $n = 2$ in equation 7: a similar approach was also taken by [74]. As we set $m/n = 0.45$ in our $k_{sn}$ calculation, this results in $m = 0.9$. We then separate the calculated erodibilities based on tectono-stratigraphic unit and calculate the median $K$ for each. The median values of $K$ for each unit can be found in Table S1 in the Supplementary Information.

Other approaches to estimating erodibility have suggested an erodibility index that incorporates i) a rock strength index ($L_L$), related to its composition, and ii) an age index based on the stratigraphic age of the unit [75, 76]. We also tested this method of determining erodibility and found that it did not alter the relative importance in the random forest analysis (Supplementary Information).
**Random forest regression.** Random forest (RF) regression is a powerful non-parametric approach to regression, which allows the determination of variable importance (VI) [38]. We performed RF regression on the 10 km gridded dataset to isolate the key signals of valley widening and reduce dataset noise. Before running the regression model we split the gridded dataset into 80% training and 20% testing to allow for validation.

The number of decision trees (\(N_T\)) used to build the regression model has shown to be important when using RF regression, particularly when investigating VI [37]. We therefore performed a sensitivity analyses on the regression varying the number of decision trees from 10 to 2000 (Extended Data Fig. 1). This analysis showed that the root mean square error (RMSE) of the regression model became relatively insensitive when the number of decision trees is greater than 1000, with RMSE 167 m. We therefore ran all RF regression runs with 1000 decision trees to ensure greatest computational efficiency.

VI in random forest regression can be determined through two approaches: average impurity reduction; and permutation reduction [e.g. 37, 79, 80]. Average impurity reduction [37] states that the importance (\(Imp\)) of any variable \(X_j\) in predicting the target variable, \(Y\), can be calculated by summing the weighted impurity decreases \(p(t)\Delta i(s_t, t)\), where \(t\) represents each node where \(X_j\) is used, and \(\varphi_m\) is tree \(m\) in the forest containing all trees \(m = 1, ..., M\):

\[
Imp(X_j) = \frac{1}{M} \sum_{m=1}^{M} \sum_{t \in \varphi_m} \delta_{j_t,j} [p(t)\Delta i(s_t, t)],
\]

(8)

where:

\[
\delta_{j_t,j} = \begin{cases} 
1 & \text{if } j_t = j \\
0 & \text{otherwise},
\end{cases}
\]

(9)

\(p(t)\) is the proportion of samples reaching \(t\), and \(j_t\) is the variable used to split node \(t\) [80]. This approach gives the most importance to the variable that most decreases the mean impurity across all trees in the forest. However, the impurity reduction approach has been shown to be biased towards predictors that have a large number of values [81]. Therefore, an alternative approach to estimating variable importance called permutation reduction has been suggested [37], which estimates the change in the mean standard error of the regression model when permuting a variable. The reader is referred to [37] and [80] for a full derivation and discussion of permutation reduction VI. We performed a sensitivity analysis of the variable importances derived for the valley-floor width regression model to choice of VI metric across a range of different decision trees (Extended Data Fig. 2). We find that the VIs are insensitive to the number of decision trees used in the regression model, and that the order of VI is identical with our chosen model run of 1,000 trees.

**Code availability.** The code for topographic analysis, including valley-floor width extraction, is available as part of the open-source LSDTopoTools software package [50]. The code to estimate exhumation rates from
Extended Data Fig. 1 The relationship between root mean square error (RMSE) of predicted valley-floor widths and number of decision trees in the random forest regression model, showing that RMSE becomes insensitive when \( \geq 1000 \) decision trees are used.

Extended Data Fig. 2 The relationship between variable importance for each predictor in the regression model and number of decision trees in the random forest regression model for (a) average impurity VI; and (b) permutation VI.

thermochronology data is available through the Zenodo data repository (https://doi.org/10.5281/zenodo.7053218).

Data availability. The thermochronometric dataset used in this paper is available through the Zenodo data repository (https://doi.org/10.5281/zenodo.7053115). Due to the large file sizes, please
contact the corresponding author if you would like all or part of the valley-floor width dataset.

**Methods References**


Himalayan valley-floor width


Himalayan valley-floor width


**Supplementary information.** This article has supplementary information.

**Acknowledgments.** We thank Steffi Tofelde, Alex Densmore, Rebecca Hodge, Mark Allen, and Elizabeth Dingle for useful discussions which helped to improve the manuscript. LSDTopoTools software development was supported by a Durham Research Development Fund grant and NERC grants NE/P012922/1 and NE/S009000/1. For the purpose of open access, the authors have applied a Creative Commons Attribution (CC BY) licence to any Author Accepted Manuscript version arising.

**Declarations**

**Author contributions.** F.J.C., S.M.M., H.D.S., and R.D. developed the study. F.J.C. and S.M.M. developed the topographic analysis code. F.J.C. performed the topographic analyses, the random forest regression, and created the figures. T.F.S. and P.A.B. compiled the thermochronometry data and performed the exhumation rate calculations. F.J.C. wrote the paper with contributions from all authors.

**Competing interests.** The authors declare no competing interests.

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Supplementary Files

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