Topographic signatures of progressive glacial landscape transformation

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Summary
The Pleistocene glaciations left a distinct topographic footprint in mountain ranges worldwide. The geometric signature of glacial topography has been quantified in various ways, but the temporal development of landscape metrics has not been traced in a landscape evolution model so far. However, such information is needed to interpret the degree of glacial imprint in terms of the integrated signal of temporal and spatial variations in erosion as a function of glacial occupation time.

We apply a surface process model for cold-climate conditions to an initially fluvial mountain range. By exploring evolving topographic patterns in model time series, we determine locations where topographic changes reach a maximum and where the initial landscape persists. The signal of glacial erosion, expressed by the overdeepening of valleys and the steepening of valley flanks, develops first at the glacier front and migrates upstream with ongoing glacial erosion. This leads to an increase of mean channel slope and its variance. Above steep flanks and head-walls, however, the observed mean channel slope remains similar to the mean channel slope of the initial fluvial topography. This leads to a characteristic turning point in the channel slope–elevation distribution above the equilibrium line altitude, where a transition from increasing to decreasing channel slope with elevation occurs. We identify this turning point and a high channel slope variance as diagnostic features to quantify glacial imprint.

Such features are abundant in glacially imprinted mid-latitude mountain ranges such as the Eastern Alps. By analysing differently glaciated parts of the mountain range, we observe a decreasing clarity of this diagnostic morphometric property with decreasing glacial occupation. However, catchments of the unglaciated eastern fringe of the Alps also feature turning points in their channel slope–elevation distributions, but in contrast to the glaciated domain, the variance of channel slope is small at all elevation levels.

KEYWORDS
Eastern Alps, erosion, glacial geomorphology, landscape evolution, numerical modelling
INTRODUCTION

Late Cenozoic climate cooling, with periods of intense glaciations in mid- and high-latitude mountain ranges, led to the formation of characteristic landforms such as knife-edged ridges, horns and cirques above the equilibrium line altitude (ELA) of the last glacial maximum (LGM) and wide U-shaped valleys with distinct overdeepenings below (e.g. Davis, 1906; Penck, 1905; Preussler et al., 2010). Such climate-specific landforms show that climaxes exert a primary control on mountain topography (e.g. Davis, 1906; Herman et al., 2013; Hinderer, 2001; Peißen et al., 2001; Penck, 1905; Robl et al., 2015b) and allow the identification of climate impact by analysing landscape geometry (Brocklehurst & Whipple, 2004; Pedersen et al., 2010; Prasicek et al., 2015; Sternai et al., 2011). Geomorphometric studies highlight that the fraction of surface area in fluvial steady-state landscapes decreases monotonically from base level to summit areas, whereas glacially imprinted mountain ranges feature hypsometric maxima at elevations between the modern and long-term glacial ELAs (e.g. Brocklehurst & Whipple, 2004; Egholm et al., 2009; Pedersen et al., 2010; Sternai et al., 2011).

However, a transient mountain landscape represents just a snapshot of its long-term evolution. The geometry of the original fluvial landscape is in general unknown, hence quantifying the impact of late Cenozoic climate change on mountain topography based on landscape metrics represents a difficult problem. This prevents the interpretation of the evolutionary state of glacial topography. Such knowledge about fluvial landscapes, however, has significantly advanced the field of tectonic geomorphology, and a glacial counterpart is needed. Only a few attempts exist to provide theory (Headley et al., 2012; Prasicek et al., 2015; Sternai et al., 2011). Geomorphometric studies highlight that the fraction of surface area in fluvial steady-state landscapes decreases monotonically from base level to summit areas, whereas glacially imprinted mountain ranges feature hypsometric maxima at elevations between the modern and long-term glacial ELAs (e.g. Brocklehurst & Whipple, 2004; Egholm et al., 2009; Pedersen et al., 2010; Sternai et al., 2011).

In combination, the ratio between the destruction of peak relief and the formation of valley relief controls whether total mountain relief decreases or increases during cold-climate periods (e.g. Robl et al., 2020). With increasing duration of glacial occupation, the combination of both processes (Figure 1c) should eventually lead to large low-relief areas above the ELA and deep valleys with steep flanks below (e.g. Egholm et al., 2017; Steer et al., 2012). This results in a topographic signal with low mean channel slopes at high elevations but high mean channel slopes at lower elevations (Robl et al., 2015a).

In the following, we systematically explore the progressive transformation from fluvial to glacial landscapes by employing a landscape evolution code capable of simulating glacial and periglacial processes (Egholm et al., 2011, 2015). Starting with a fluvial landscape in steady state, we follow the trajectory of landscape change and determine loci and points in time where topographic changes reach a maximum (hot spots) and where the initial landscape persists (cold spots). We trace the fluvial-to-glacial transformation of an entire mountain range and a single valley, and compare derived landscape metrics from model time series with real-world examples of mid-latitude mountain ranges such as the Eastern Alps.
2.1 | Data

Morphometric analyses of natural landscapes are based on the EU-DEM (data funded under GMES, Global Monitoring for Environment and Security preparatory action 2009 on Reference Data Access by the European Commission) with a spatial resolution of approximately 25 m. For time-dependent numerical models, where natural landscapes represent the initial state, we use a bilinear interpolation method to resample gridded elevation data to resolution of \( \frac{1}{24} \) th

2.2 | Distribution of channel slope versus surface elevation

We analyse the distribution of channel slope versus surface elevation, an approach that has been applied to both fluvial and glacial terrain (Hergarten et al., 2010; Künni & Pfiffner, 2001; Robl et al., 2015a). Channel slopes in m/m are calculated by a standard single flow algorithm (D8) implemented in Topotoolbox (Schwanghart & Kuhn, 2010; Schwanghart & Scherler, 2014). All cells defined by a flow accumulation greater than one pixel in the synthetic example and four pixels in the natural landscapes (>0.04 km²) are considered for computing channel slopes. We classify the channel slope data into 50-m-wide elevation bins and report statistical indicators such as mean and percentiles (P25, P50 and P75) for each bin. For a graphical representation of the channel slope distribution per elevation bin, the relative frequency of the occurrence of channel slope values is evaluated using channel slope bins with a width of 0.02 m/m (Robl et al., 2015a). Further, we investigate the shape of the entire channel slope–elevation distribution.

2.3 | Landscape evolution model

The computational landscape evolution model employed here can simulate glacial, fluvial and hillslope erosion. The fluvial component is only used here to provide a fluvial steady-state topography as a reference and starting point to systematically explore the fluvial-to-glacial landscape transformation. To obtain a fluvial steady-state landscape, we employ a detachment-limited model for bedrock channel incision (e.g. Braun & Willett, 2013; Hergarten, 2002; Howard, 1994, 1980). Erosion in the hillslope domain is approximated by diffusion to produce the initial fluvial steady state.

In the glacial model, the flow of ice is handled by an integrated second-order shallow ice approximation scheme (iSOSIA) (Egholm et al., 2011). This approach aims to balance the requirements of accuracy in simulating the ice flow and computational efficiency in modelling the long-term landscape evolution. In contrast to standard shallow ice approximation (SIA) models, iSOSIA considers longitudinal and transverse stresses in the ice and hence resolves horizontal stress gradients, which are important for alpine settings, because they provide local coupling of ice flow and impede strong local gradients in ice velocity (Egholm et al., 2012b). Accumulation and ablation of ice are modelled by a simple positive-degree-day approach, and additionally modified by avalanches and snow drift (Egholm et al., 2015). We model ice temperate and ignore the effects of basal freezing at the upper reaches of thin, high-altitude alpine glaciers. Glacial erosion processes are represented by abrasion and quarrying. We model glacial erosion on bedrock and neglect the potential effects of subglacial sediment protecting the glacier bed. Abrasion is estimated based on ice sliding at the glacier base (Egholm et al., 2012a; Hallet, 1979; Lilbreit, 1994). We use a sliding law that accounts for the opening of cavities due to the roughness of the bed (Ugelvig et al., 2018b). Consequently, subglacial hydrology controls the effective pressure of
the system and influences the basal sliding speed (Ugelvig et al., 2016). Subglacial quarrying is additionally influenced by effective pressure directly and bed channel slope in the direction of sliding (Iverson, 2012; Ugelvig et al., 2016).

In the glacial model, a temperature-dependent frost cracking approach is employed to model periglacial erosion (Andersen et al., 2015; Egholm et al., 2015). Since steep surface slopes dramatically reduce the performance of the numerical model, a local slope threshold criterion is used. This criterion immediately lets all local slopes steeper than a threshold of 1 (45°) collapse to the threshold slope. While this criterion is required for numerical reasons, it can be interpreted as some kind of landsliding.

Tectonic uplift is only considered to produce the initial fluvial topography, while isostatic effects related to loading/unloading of the lithosphere due to ice and erosion are also included. Detailed description of the combined model can be found elsewhere (Braedstrup et al., 2016; Egholm et al., 2011, 2012a, 2015; Ugelvig & Egholm, 2018a). However, a summary of the setup and chosen parameter values can be found in the Supplementary Information. We use standard parameter sets for all model components. In the following, we briefly describe the most important aspects of the model setup.

3 | GENERAL MODEL SETUP

The initial topography for all the synthetic experiments has an extent of 60 × 80 km and consists of 10,000 irregular Voronoi cells. We start with a 500-m-elevated plain and some low-amplitude random noise. While the model domain is subsequently uplifted, the initial base level of 500 m at the model boundaries is preserved. We choose this base level to avoid bedrock elevations below sea level in regions of major glacial overdeepening. For all simulations, we uniformly set a fluvial erosion constant (K_f) of $1 \times 10^{-5} \text{a}^{-1}$, and the stream power exponents n_1 and n_2 to 0.5 and 1, respectively. Based on these initial and boundary conditions, we apply (a) a tent-shaped uplift pattern with zero uplift at the southern and northern boundary and a maximum uplift rate of 0.002 m/a at the center of the range (see Figures 2–7) and (b) a uniform uplift pattern with 0.002 m/a within the model domain and 0 m/a at the boundaries (see Supplementary Figures S4–S9). Both simulations are performed until a balance between uplift and erosion is met at almost every point of the grid and topographic steady state is nearly reached. All relevant model components and parameters are summarised in Supplementary Tables S1 and S2.

Before introducing cold surface processes, the fluvial topography is interpolated onto a regular grid with 300 × 400 cells (i.e., 200 m resolution) to reduce computation time. For both experiments (a) and (b), we set a constant sea level temperature of 6.5°C and 8.5°C, respectively, a lapse rate of 6°C/km, a seasonal temperature variation of 14°C and a positive-degree-day melt rate constant of 0.001 m d⁻¹ °C⁻¹, which results in an ELA of ~1300 m and ~1650 m. Precipitation is set to 1 m/a for the entire model domain. These mass-balance parameters are chosen to cover the landscape with an extended ice layer so that the initial ice accumulation areas in both experiments have roughly the same size. The free-scaling erosion parameters K_q and K_s are uniform in all our experiments as we are interested in glacial erosion patterns controlled by pre-glacial topography and not on the impact of substrate properties (i.e., lithology). They are set to $1 \times 10^{-3} \text{Pa}^{-1}$ and $1 \times 10^{-4} \text{Pa}^{-1}$, respectively.

Our model setup provides little capacity for frost cracking, because a constant mean annual temperature limits the landscape to be exposed to frequent freeze–thaw cycles (Anderson, 1998; Hales & Roering, 2007). In addition, a thick ice cover dampens the surface temperature variations needed for promoting segregation ice growth.

To account for the glacial–interglacial cycles with waxing and waning glaciers typically observed in glaciated mountain ranges at mid-latitudes, we perform a third experiment (c) where we introduce three asymmetric, sawtooth-shaped temperature cycles with a period of 100 ka and an amplitude of ±3°C. Each cycle is represented by 80 ka of constant cooling followed by 20 ka of rapid warming, which results in an ELA varying between ~1300 and 1850 m. We stress that this approach is not meant to simulate a specific time period but simply to account for general temperature variations in the long-term climate record (i.e., Lisiecki & Raymo, 2005) and their specific effects on topography (e.g., cirque formation).

4 | RESULTS

In the following, we explore the emerging spatial and temporal patterns and changes in landscape geometry. First, we highlight topo-graphic changes due to glacial and periglacial erosion on a mountain range scale. Second, we track temporal and spatial variations in glacial erosion and the resulting topographic patterns in a single catchment. Finally, we compare the model results with selected catchments of the Eastern Alps to estimate the degree of glacial reshaping.

4.1 | Spatio-temporal pattern of glacial erosion on a mountain range scale

The initial fluvial mountain range is dominated by a west–east-trending principal drainage divide with a maximum elevation of ~2150 m (tent-shaped uplift) and shows a maximum relief of ~1650 m (Figure 2a). Major valleys originate at the principal divide and mostly follow the uplift rate gradient towards the northern and southern boundary. Trunk valleys and their mostly west–east- and east–west-draining tributaries are V-shaped in cross-section and characterized by graded longitudinal profiles.

After 300 ka of glacial erosion, the main landscape elements such as the central ridge and the major north- and south-draining valleys persist (Figure 2b). However, originally V-shaped valleys become distinctly deeper and wider and form characteristic glacial troughs with U-shaped cross-sections and overdeepenings as major deviations from their initially graded longitudinal profiles. Smaller tributary catchments become hanging valleys with distinct elevation drops at the confluence with the trunk valley. Such distinct changes in the appearance of the landscape result from strong spatial differences in glacial erosion. While total erosion exceeds 500 m in trunk valleys where overdeepenings are formed, most tributaries show less than 100 m of deepening and ridges remain almost unaffected by erosion (Figure 2c). This spatial variability of erosion causes a strong increase in local relief, which is primarily controlled by the accumulation and flux of ice and the availability of water at the glacier base (Figure 2d).
Low-order drainages below the ELA, ridge lines with negligible ice accumulation area and steep valley flanks prone to avalanching remain almost ice free. In contrast, all trunk valleys show a maximum ice thickness of 500 m and more, which is thinning towards the glacier front and upstream in the tributaries. As the glacier mass balance and hence ice flux and thickness are primarily controlled by topography, a feedback arises between glacial erosion and accumulation of ice. This cannot be explored by analysing topographic end-members (i.e. full fluvial, Figure 2a; full glacial, Figure 2b) but requires analysing time series showing the progressive landscape transformation (Figure 3).

Therefore, we extract time series of a swath profile perpendicular to the principal divide and of three profiles paralleling the central crest of the mountain range (see Figure 2a, b for the position of the profiles). Before glaciation, the mountain range becomes increasingly steeper with elevation, following an increase in uplift rate towards the central crest (see the conceptual Figure 1 for comparison). Glacial erosion alters the topography only within the domain of glacial occupation, which is limited to about 20 km north and south of the main ridge. Despite 300 ka of glacial reshaping, differences in mean and maximum elevation between the initial and the glacially reshaped landscape are small. In contrast, major differences are seen in the minimum elevation representing valley floors, with the greatest amount of surface lowering (up to 500 m) at the central part of major valleys, where our model shows the formation of overdeepenings (Figure 3a).

At the lower reaches (Figure 2, profile B), most of the surrounding landscape lies below the ELA, and large parts of the fluvial characteristics of the landscape persist. Topographic changes are limited to the main trunk of major glaciated valleys, which are progressively widened and lowered by about 200–400 m (Figure 3b). In one valley, located roughly at kilometre 20 along the transverse profile (Figure 3b), the valley floor experiences a distinct boost in erosion. From 100 to 300 ka, a maximum of 600 m of rock is eroded. The fluvial-to-glacial topographic evolution of this catchment is described in detail in Sect. 5.

Further upstream (Figure 2, profile C), the highest ridges and peaks protrude the regional ELA of ~1300 m (Figure 3c). Erosion
focuses close to the main valley trunk, but in addition, small-scale ridges beneath the ELA and close to the trunk of large valleys are overflown by ice and progressively removed, which results in valley widening (Figure 3c) and valley straightening (Figure 2).

The headwater section (Figure 2, profile D) near the central ridge is entirely located above the ELA (Figure 3d). However, apart from a few exceptions, the fluvial topography persists at the summit domain, which is characterized by ice-free conditions, or covered by a very thin ice layer due to steep hillslopes and avalanching. Minor topographic changes within the headwaters are primarily caused by headwall erosion. Ice-free regions are affected by periglacial processes (i.e. temperature-dependent frost cracking). However, these processes result in just a few metres of erosion, which is hardly visible on the scale of the profiles.

The western and eastern fringe of the mountain range are affected by strong glacial erosion (see also Figure 2 for the spatial distribution of glaciers). In particular, valleys in the eastern part of the mountain range are lowered by several hundred metres with an erosion hot spot at a distinct overdeepening, where the total erosion exceeds 500 m. This observation is not surprising as it stems directly from the prescribed accumulation pattern of ice, which limits ice accumulation at steep summit domains and promotes ice ablation, and hence erosion, in trunk valleys.

4.2 | Characteristics of landscape change in the channel slope–elevation domain

As the ELA position and the concentration of ice flux along the drainage lines control the timing and spatial pattern of glacial erosion, our results reveal that the transformation from a fluvial to a glacial landscape affects different elevation levels of alpine catchments to varying degrees. This is reflected in the distribution of channel slope with elevation. In particular, temporal changes of the channel slope–elevation distribution are diagnostic for the degree of fluvial-to-glacial landscape transformation.

The initial fluvial landscape shows a characteristic channel slope–elevation distribution for an uplift pattern with increasing uplift rates from the peripheral to the central parts of the mountain range (tent-shaped uplift pattern). The mean channel slope increases continuously with elevation (Figure 4a) so that an increase in channel steepness corresponds to an increase in uplift rate (see also Figure 1, dashed line for comparison). For the given set of model parameters, we observe a narrow frequency distribution of channel slopes at all elevations with maximum channel slopes below 1. The variations in channel slope within an elevation interval result from the distribution of the contributing drainage area of the channel segments included in the elevation interval and from the fact that areas of equal height in a mountain range do not necessarily coincide with areas of equal uplift rate.

Once cold-climate surface processes start reshaping fluvial topography, the mean and variance of the channel slope increase above the ELA (Figure 4b). This trend is amplified with ongoing glacial occupation. After 300 ka of glacial erosion, the topography is represented by a channel slope–elevation distribution with a distinct channel slope maximum slightly above the ELA and a second maximum below the initial base level where glacial troughs formed (Figure 4c, d). The spread between the first and third channel slope quartile distinctly increases from the initial fluvial to final glacial landscape across all elevations, with the strongest rise at and above the ELA, around 1500–1800 m of altitude.
At \( \sim 1650 \) m, the frequency of high channel slopes rises from \( \sim 25\% \) in the fluvial to \( \sim 75\% \) in the glacial landscape. After 300 ka of glacial erosion, a second maximum of high channel slopes emerges below the initial base level due to the formation of glacial overdeepenings (Figure 4d, grey bars). Hence, both the general channel slope–elevation distribution and the fraction of high channel slopes are possible diagnostic metrics to constrain the degree of glacial reshaping of a fluvial landscape.

We assume that the synthetic landscape approaches a characteristic glacial topography after 300 ka of continuous ice occupation, as conceptualized in Figure 1. However, as the glacier mass balance, and hence the erosional potential, are primarily controlled by climate (i.e. ELA position) and topography, we also investigate the influence of climate fluctuations (Supplementary Figs. S1–S3) and of an initial fluvial landscape geometry representing uniform instead of tent-shaped uplift (Supplementary Figs. S4–S9).

Starting with the same initial landscape as shown in Figure 2a but imposing 100-ka climate cycles with an ELA fluctuating between 1300 m and 1850 m, we observe similar temporal changes in mean channel slopes when compared with our experiment with constant climate conditions (Supplementary Fig. S3). However, due to the reduced glacial occupation time, the glacial imprint is generally less pronounced. This applies in particular to the lower reaches of glaciated catchments, where the formation of glacial overdeepenings is less pronounced (Supplementary Figs. S1 and S2). Consequently, we do not observe a maximum of high channel slopes emerging below the initial base level as observed in experiments assuming constant cold-climate forcing (compare Figure 4d and Supplementary Fig. S3).

The channel slope–elevation analysis reveals the glacial impact on topography on the scale of the orogen. To identify locations and timing of topographic changes, we track the development of erosional hot spots and cold spots over time in a single catchment. This also facilitates the comparison with observations in real-world examples. The initial catchment topography shows several bends in the main ridge line indicates that large parts of the landscape remained unaffected by glacial erosion processes and abundantly show even after 300 ka of glacial erosion largely initial landscape metrics (Supplementary Fig. S1c). As consequence of a fluctuating ELA, low-gradient valley floors (i.e. cirques) are shaped at various altitudes, but these areas are rather small and have a minor effect on channel slopes.

Assuming uniform uplift, the initial fluvial mountain range features a peak elevation of 2280 m and a valley network topology systematically deviating from a mountain range with a tent-shaped uplift pattern. This initial landscape configuration results in a distinctly different ice configuration with characteristic spatial and temporal patterns of glacial erosion (Supplementary Figs. S4–S9) eventually reflected in the distribution of channel slope with elevation. The initial topography shows a strong increase in mean channel slope at low elevations and flattens towards a value of 0.6 with increasing elevation (Supplementary Fig. S6). With ongoing glacial occupation, we observe a small but steady increase of mean and variance of channel slope around the ELA (Supplementary Fig. S6). Compared with previous experiments, the temporal changes in mean channel slope are rather small as the ice flow pattern consists of a large number of small glaciers forming less local relief. The numerous interfluves remain largely unaffected by glacial erosion, which preserve the initial fluvial landscape characteristics (Supplementary Fig. S4c).

4.3 Evolution of topographic changes in a single glacier catchment

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Major topographic modification occurs in the elevation range between the minimum and maximum ELAs representing interglacial and full glacial periods. Within this elevation range, we observe a distinct increase in channel slope and its variance (Supplementary Fig. S3). Similar to constant climate conditions, the observed increase in mean channel slope at mid-altitudes is related to the formation of steep valley flanks, whereas the decrease of channel slopes towards the ridge line indicates that large parts of the landscape remained unaffected by glacial erosion processes and abundantly show even after 300 ka of glacial erosion largely initial landscape metrics (Supplementary Fig. S1c). As consequence of a fluctuating ELA, low-gradient valley floors (i.e. cirques) are shaped at various altitudes, but these areas are rather small and have a minor effect on channel slopes.

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fluvial valley, which form interlocking spurs (Figure 5a). Along the trunk valley, glacial erosion efficiently removes these spurs and straightens the path of ice flow. Flattened longitudinal valley segments emerge, confined by steep valley flanks and terminated by small cirques and headwalls. In addition to the excavation of an extended glacial trough, hanging valleys form where tributaries enter the main trunk (Figure 5b, arrows).

In general, the pattern of glacial erosion shows strong spatial variations. This causes a significant increase in local relief from the original fluvial to the resulting glacial landscape (Figure 5e and f, respectively). While the fluvial landscape shows smooth variation in local relief with a slight increase towards the headwaters, high values of local relief highlight erosion hot spots occurring during the fluvial-to-glacial landscape transformation. We observe a distinct increase in relief at the central section of the trunk valley and at the headwaters close to the mountain crest where local relief exceeds 750 m (Figure 5e, f).

The progress of the glacial landscape transformation over time is highlighted by the time series of catchment-wide swath profiles in Figure 6a. At the lower reach, the difference in elevation shows a maximum between the initial fluvial and the resulting glacial valley length geometry. There, a major overdeepening of several hundred metres formed within the ablation zone of the glacier trunk, far below the ELA (Figure 6a). On average about 500 m of rocks are eroded at this valley section.

The longitudinal channel profile of the initial fluvial trunk valley (roughly expressed by the minimum elevation of the swath profile in Figure 6a) appears smooth and is well graded with a steep upper reach in the headwaters and a decrease in channel gradient with an increasing contributing drainage area in the downstream direction. In contrast, the glacially reshaped valley has a wide, low-gradient valley floor punctuated by multiple steps and overdeepening ~350 m below the initial base level of 500 m. Further upstream, local small-scale topographic depressions vanish as the valley floor is uniformly lowered by 100–150 m relative to the initial fluvial profile. Hence, channel gradients do not significantly change with the transition from fluvial to glacial conditions in this valley segment. However, as the ridge-line elevation remains fairly constant, the glacial longitudinal profile steepens more rapidly towards its headwaters than in the fluvial case. This results in the observed local relief increase within the headwaters domain (Figures 5 and 6d).

In our simulation, the valley trunk is excavated faster and deeper than its tributaries and the surrounding topography. This is shown by the time series of cross-sections perpendicular to the main ice flow in Figure 6. From the lower trunk up to where the main ice stream bifurcates (Figure 5 profile C’ and B’), glacial erosion substantially lowers and widens the trunk valley, while adjacent domains remain unaffected and fluvial characteristics persist (Figure 6b, c). Further upstream (Figure 5 profile C), the ice cover is more extensive and acts

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**Figure 5**  Topographic evolution of a single initially fluvial catchment under a constant cold climate (a–f). Position of catchment is shown in Figure 2a and b (black outline). Red lines and symbols display the positions of profiles shown in Figure 6 and locations shown in Figure 7, respectively. Numbered arrows point to (1) a truncated spur, (2) a flattened valley floor, (3) a cirque, and (4) a hanging valley. Surface elevation contour spacing in (d) is 100 m. ELA contour line (~1300 m) is shown in white.

**Figure 6**  Changes in longitudinal and transverse valley geometry due to glacial erosion. (a) Time series of maximum, mean and minimum elevations of the longitudinal swath profile. (b–d) Time series of transverse profiles.
as an erosive agent not only at the main trunk, but also along its tributaries by variably lowering the entire landscape and removing small-scale topographic features such as ridges (Figure 6d).

By comparing the topographic changes between each 100-ka-long time interval, we observe that the associated valley widening occurs at an early stage of glacial reshaping, while the rate of valley incision amplifies with the duration of glacial occupation and progressive valley overdeepening (Figure 6a–c). Time series of topographic changes along the different profiles in the catchment indicate that overdeepening starts at the lower part of the main trunk and propagates progressively in the upstream direction (Figure 6b, c).

By tracing the evolution of topography at characteristic points, our model reveals the spatial and temporal changes in glacial erosion (Figure 7). We track the evolution of the land surface at three elevation markers (Figure 5a, b) located at the ridge line (square), the valley flank (triangle) and the valley floor (circle) through time. The three locations experience a distinctly different erosion history (Figure 7a). While the ridge line elevation does not change, the valley floor elevation is progressively lowered, and the erosion accelerates during overdeepening of the major valley. The valley flank elevation remains constant for more than 250 ka of glacial occupation, but eventually drops as a consequence of progressive valley widening (Figure 7a, blue triangles).

The valley floor is lowered by several hundred metres at rates that vary in space and time (Figure 7b). While the valley floor at the upper valley section (black circles) experiences only moderate erosion of less than 100 m and shows only a slight increase in rate, valley floors at the middle (blue circles), lower-middle (magenta circles) and lower sections (red circles) are affected by a total erosion of several hundred metres. With the onset of glacial occupation, erosion rates are highest at the lower valley section and further increase from ~0.001 m/a to ~0.003 m/a after about 50 ka of glacial erosion. After about 210 ka, the erosion rate tapers to its initial value (Figure 7b). The lower-middle and middle valley sections experience an even faster increase in valley floor erosion at about 220 ka and 260 ka, respectively. There, total erosion eventually approaches a similar total erosion as in the lower valley section.

Glacial incision into the valley floor changes the relief of the valley flanks. With the excavation of the trunk valley and flank elevations remaining fairly constant, the cross-sectional relief of the trunk valley increases with time (Figure 7c). This kind of relief formation accelerates with the overdeepening of the lower and middle valley segment, while the relief at the upper valley section declines for ~220 ka as valley flanks are lowered more efficiently due to headwall erosion than the adjacent floor (Figure 7c, black triangles). However, the cross-sectional relief increases as soon as the erosive signal of the main trunk has propagated towards the upstream section (Figure 7b, black circles) and eventually approaches the initial relief (Figure 7c, black triangles).

**Figure 7** Evolution of surface elevation and relief at characteristic locations of the catchment. Locations are shown in Figure 5a and b. Circles represent the valley floor, triangles the valley flanks and squares the ridges. Colours indicate the upstream distance along the valley. Red, magenta, blue and black indicate 15 km, 17.5 km, 20 km and 25 km, respectively. (a) Temporal evolution of ridge, valley flank and valley floor elevation along the transverse profile at 20 km. (b) Evolution of the valley floor elevation at the lower, lower-middle, middle and upper valley section. (c) Evolution of the valley relief at the lower, middle and upper valley section. The relief is computed by subtracting the elevation of the valley floor from the elevation of the flank along the same transverse profiles. (d) Evolution of local relief (radius 500 m) at the valley floor. (e) Same as (d) but overdeepenings are filled. Note that the first uniform drop in elevation of about 20 m at the very beginning of the model run reflects the isostatic response due to ice loading. (f) Evolution of valley overdeepening. Volume (black stars) and area (red squares) of overdeepenings are calculated by subtracting the elevation of the modelled topography from a topography with synthetically filled depressions. Depressions are defined by a minimum depth of 20 m. (g) Double-logarithmic area-volume relation plot. Gradients 1 and 1.5 are shown for comparison.
In our simulation, glacial erosion processes widen and straighten the valley due to the rapid destruction of morphological obstacles. During the first 80 ka, the local relief remains nearly constant at the lower section. Thereafter, local relief starts to increase with the formation of an overdeepening (Figure 7d). At 220 ka, local relief reaches a maximum of 550 m and decreases again thereafter due to progressive valley widening. Local relief shows a similar development at the middle valley section, starting at about 160 ka. It increases until 250 ka towards a maximum value of 400 m, then stagnates thereafter. At the upper section, local relief decreases with glacial erosion for 230 ka then increases slightly during the remaining 70 ka.

Overdeepenings in natural landscapes are immediately filled by water and sediment. To account for this, we filled depressions before computing local relief (Figure 7e). This results in a valley floor relief of (nearly) zero at the lower and middle reach (Figure 7e, red and blue circles). At the upper part of the valley, only minor overdeepenings form and filling of depressions has only a minor impact on the valley floor relief (Figure 7e, black circles).

The evolution of the valley overdeepening is also strongly influenced by spatial and temporal changes of glacial erosion. At the beginning of landscape transformation, both the area and volume of valley depressions show only a small linear increase (Figure 7f). At this stage, the topography is characterized by small patches of valley depressions, predominately located at the glacier terminus and valley heads (Supporting Information Movie S2). Associated with the strong increase of erosion at the trunk valley floor (Figure 7b, magenta and blue circles), a prominent overdeepened glacial trough starts to form at about 200 ka (Figure 7f). We observe this as a strong rise in the volume of valley depressions compared with their area expressed as a gradient greater than 1.5 in the double-logarithmic area-volume relation plot (Figure 7g, black line).

4.4 The signature of glacial erosion in three alpine catchments

We compare the progressively evolving glacial signature from our modelled synthetic landscapes with landscape characteristics from catchments of the Eastern Alps to determine the degree of glacial imprint in mid-latitude mountain ranges. In the Eastern Alps, fully, partly and non-glaciated catchments with similar peak elevation, substrate properties and structural inventory exist in spatial proximity (Figure 8a). This allows elimination of the otherwise strong lithological control on mountain topography (i.e. the persistence of fluvial and glacial landforms) (e.g. Prasicek et al., 2015; Robl et al., 2015a).

We choose the Trofaiach catchment (Mur) (Figure 8c), the Bretstein catchment (Mur) (Figure 8d) and the Flachau catchment (Enns) (Figure 8e) as characteristic representatives of the minorly, partly and fully glaciated domain of the Eastern Alps, respectively. While the Bretstein catchment is situated at the edge of the LGM ice extent, only minor valley glaciers existed in the Trofaiach catchment throughout the Quaternary (e.g. Ehlers et al., 2011) (Figure 8a). Longitudinal channel profiles reveal morphological differences in each catchment. The main stem of the Trofaiach catchment is well graded with two distinct knickpoints. In contrast, the Bretstein and Flachau catchments are characterized by low-gradient trunk channels with several smaller knickpoints in their lower courses, and a strong increase in the channel gradient towards their headwaters, which is more pronounced in the formerly fully glaciated Flachau catchment (Figure 8b). In the glaciated domains, the tributaries form hanging valleys.

The occurrence of glacial landforms with low-gradient valley and cirque floors in direct vicinity to steep flanks, cirque walls, ridges and horns is accompanied by differences in the channel slope–elevation distributions of the fluvial and glacial catchments. The channel slope–elevation distribution of the Trofaiach catchment,
represents predominantly fluvial conditions, shows a monotonic increase in the mean channel slope with elevation (Figure 8f). In contrast, the partly glaciated Bretstein catchment shows a mean channel slope increase up to an elevation of about 1450 m followed by stagnating mean channel slopes and a second increase at 1800 m (Figure 8g). The fully glaciated Flachau catchment shows a mean channel slope increase up to an elevation of about 2000 m, and a decrease from about 0.6 at (tributary) headwalls to about 0.35 at the summit domains (Figure 8h). The channel slope–elevation distributions of the Trofaiach and Bretstein catchments show a smaller interquartile range in channel slope compared with the fully glaciated Flachau catchment. The latter shows a scattered channel slope distribution in all elevation ranges.

Applying our cold surface process model with the same model parameters (i.e. ELA ~1300 m) to the Trofaiach catchment leads to glacial overprinting similar to model runs on synthetic landscapes and as observed in natural catchments. The former narrow distribution of channel slopes becomes scattered over a wide range of channel slopes at all elevation levels (Figure 9b). A bimodal landscape with the co-occurrence of high- and low-channel slope surfaces, expressed as valley floors and steep valley flanks, emerges (Figure 9a). We observe an overall increase in mean channel slope compared with the initial topography (Figure 9b). The evolution of mean channel slopes with elevation is now characterized by a turning point located at high altitudes where (tributary) headwalls transition to ridges and summits (Figure 9b). This transition from a monotonic increase with elevation towards a turning point of the mean channel slopes becomes more pronounced with increasing duration of glacial occupation (Figure 9b, time series from light grey to black lines). After 200 ka of glacial occupation, a similar bow shape as the mean channel slope–elevation distribution of the formerly glaciated Flachau catchment emerges (Figures 9b and 8h).

5 | DISCUSSION

5.1 | General characteristics of glacial erosion

The characteristic geometry of glacial landforms such as U-shaped valleys, bowl-shaped cirques, hanging valleys and truncated spurs indicates that glaciers are of first-order importance to landscape evolution and are reproduced by our model. The simulation of ice flow and glacial erosion and the comparison of the resulting topography with existing landscapes improve our understanding of the development of glacial landscapes (e.g. Anderson et al., 2006; Braedstrup et al., 2016; Egholm et al., 2011; Harbor et al., 1988). Here, we discuss the spatial and temporal patterns of glacial erosion emerging from our numerical experiments, the characteristics of the resulting synthetic landscapes and channel slope–elevation plots to diagnose the evolutionary state of real-world topography.

5.1.1 | Spatial variations in erosion rate

The spatial pattern of glacial erosion is controlled by climate, i.e. the ELA and the mass-balance gradient with elevation, and the initial fluvial landscape. However, the combined influence of these factors can cause an increase (Montgomery, 2002; Shuster et al., 2011; Valla et al., 2011) or decrease (Brozovic et al., 1997; Mitchell & Montgomery, 2006) of relief. Both effects have been observed in high-latitude landscapes, with an increase of fjord relief and a decrease of interfluve relief above (Egholm et al., 2017). Our simulations inspired by mid-latitude mountain ranges dominated by alpine-style glaciations with peaks and ridges towering above a warm-based glacier network show a general increase in relief.

We observe selective glacial erosion where fluvially predefined channels are excavated faster and deeper than tributaries and the surrounding topography (Figures 6 and 3). In our simulations, this results from the spatial distribution of ice flux, basal water availability and basal glacier velocity. Domains along the main channels undergo a strong fluvial-to-glacial transformation, while areas directly above with low ice flux maintain their fluvial topographic signature and hanging sections form at tributary intersections.

The vertical incision in valleys causes lengthening of flanks and headwalls and eventually forms persisting peaks that are surrounded by steep surfaces, which promote snow avalanching and reduce glacial erosion (Ward et al., 2012). However, if these ice-free flanks are within the elevation range to enable frost cracking, more rapid headwall retreat by undercutting should be possible, as the glacier below would transport the debris away and prevent the formation of
The widening and deepening of valleys in concert with the preservation of summit domains decrease the mean elevation and reduce the topographic load of mountain ranges. Isostatic compensation of localized erosion in valleys can enhance this effect and raise surrounding peaks (Robl et al., 2020).

5.1.2 Temporal variations in erosion rate

The rates of topographic adjustment to glacial conditions vary in space and time. With ongoing glacial occupation, erosional hotspots with erosion rates distinctly exceeding background rates evolve and progressively form overdeepenings, which grow both in depth and laterally. Glacial erosion rates have been constrained in general (e.g. Cook et al., 2020), but temporal variations in erosion rate showing the development and migration of erosional hotspots have only been documented in rare cases of extraordinary deep glacial valley incision (Haeuselmann et al., 2007; Shuster et al., 2011; Valla et al., 2011). For most parts of glacial landscapes, temporal changes in erosion rate remain hardly constrained. Our numerical experiments contribute to solving this issue.

As the complexity of glacial erosion processes is poorly understood, glacial landscape evolution models generally use simple erosion rules that relate the efficiency of erosion to sliding at the glacier base (Hallet, 1979; Harbor, 1992; Herman et al., 2011). Besides all its simplifications and limitations, such models allow constraining the timing and rate of glacial landscape adjustment (e.g. Pedersen & Egholm, 2013; Sternai et al., 2013). In our experiments, quarrying and abrasion represent the primary mechanisms for subglacial erosion, whereby basal sliding speed and hence the related basal shear stress are primary factors of their efficiency and control the spatio-temporal pattern of topographic adjustment from a fluvial to glacial landscape. Our experiments along with former numerical modelling studies demonstrate that subglacial shear stress decreases as glacial erosion progressively transforms preglacial V-shaped valleys into U-shaped troughs (Braedstrup et al., 2016). This is also consistent with our findings showing that the rate of channel slope adjustment is distinctly higher in the first 100 ka of glacial occupation than in the last 200 ka (Figure 4e and Supporting Information Movie S2). As a result, mean erosion rates are highest in the early stages of glacial landscape development, when the topography is not yet adapted to the new cold-climate conditions (Harbor, 1992). Spatial variations in erosion rate and the timing of maximum rates are dominated by valley overdeepening (Figures 3 and 7). Overdeepening first occurs in down-glacier reaches and progressively extends upstream (Figure 7b). We observe a strong positive feedback between the formation of overdeepenings and the sliding rate, indicating that valley excavation by glaciers is a non-linear process. Once overdeepening is initiated, the valley floor is rapidly lowered (Figure 7b, blue, magenta and red curve). Glacial hydrology plays a key role in this behaviour, as it affects subglacial sliding and thereby influences long-term landscape evolution. The highest volumes of meltwater are available in the lower part of the glacial system, increasing sliding rates and leading to a boost in erosion (Herman et al., 2011). Lowering of the glacier bed also lowers the glacier surface and hence leads to increased melting, again increasing sliding. Starting at the terminus, this signal of glacial erosion migrates upwards while the glacier progressively eats away its own base (Figure 7b) and retreats. This mechanism has the double effect of decreasing erosion rates at low elevations and triggering headward propagation of glacial erosion, which has also been shown by former glacial modelling (Sternai et al., 2013) and erosion studies (Shuster et al., 2011). Therefore, our results suggest that glacial valley deepening and erosion at and above the ELA are related processes.

Assuming constant cold-climate conditions, our model predicts only limited headwall erosion triggered by cirque glaciers incising towards the centre of the main ridge line (Figures 2c and 3d). Therefore, most of the topography at the summit domains remains unaffected by glacial erosion processes, and we do not observe a considerable development of flat cirque floors around and above the ELA. As cirque headwalls form at the highest parts of the glacial accumulation zone and our experiments model erosion via sliding-based rules, the model results indicate that downstream valley incision is far more effective than erosion of the cirque walls and floors (Figure 3). Cirques tend to experience maximal erosion during the onset and termination of glacial cycles, but limited erosion during ‘full’ glacial conditions (e.g. Alley et al., 2019; Barr et al., 2019; Evans, 2021; Yanites & Ehlers, 2012).

Introducing climate oscillations governing advancing and retreating glaciers causes a lower magnitude of glacial erosion, in particular at the lower reaches, compared with constant-climate conditions, but patterns of erosion remain similar. Again, most of the topography at the summit domains is largely unaffected (Supplementary Figs. S1c and S2d). While low-gradient valley floors sporadically evolve within the elevation range of the oscillating ELA, the size and number of such ‘cirque-like’ landforms are too small to significantly alter the channel slope-elevation distribution (Supplementary Fig. S3).

Further, headwall retreat occurs to a large part above the ice, where periglacial processes dominate erosion. This was predicted by numerical modelling (e.g. Anderson et al., 2013; Egholm et al., 2015; Hales & Roering, 2007), and shown by cosmogenic nuclide (Mair et al., 2020) and monitoring studies (Hartmeyer et al., 2020). Thus, the few metres of periglacial erosion predicted in our experiments might strongly underestimate real-world headwall retreat, at least in temperate mountain ranges. Further, the hypsometry of the initial landscape controls the mass balance and the spatial distribution of the glacial erosion potential. In a steady-state fluvial mountain range, this potential is small at highest elevations consisting of steep ridges and peaks, while it can be large in a premature landscape with elevated plateaus (e.g. Pedersen et al., 2014).

5.2 Channel slope–elevation distributions for interpreting the degree of glacial imprint

In our experiments, the glacial landscape geometry is finally characterized by spacious, flat valleys separated from persisting summit areas...
by (high) steep flanks (Figures 6 and 3). The development of such neighbouring erosional ‘hot spots’ and ‘cold spots’ creates a characteristic signature in the evolving glacially conditioned mountain range. Such a signature is visible in the channel slope–elevation distributions of the synthetic and the investigated natural mountain landscapes. In particular, it is expressed by (Figure 4e):

1. A channel slope maximum at the elevation where the valley flanks of the newly formed overdeepenings dominate the channel slope distribution.
2. A decrease in channel slope towards fluvial conditions as the largest amount of topography is not affected by glacial erosion (except major valleys).
3. A second distinct increase in channel slope roughly at the ELA with highest mean channel slopes where tributary head walls dominate.
4. A decrease in channel slope towards fluvial conditions at summit domains where the low ice thickness does not suffice for considerable ice flow and hence erosion.

These characteristics progressively intensify with the extent and duration of glacial occupation. Variations in the degree of glacial imprint can be distinguished. Thus, channel slope–elevation plots serve as a diagnostic tool for interpreting the glacial–fluvial influence in mountain landscapes.

Our modelling results of the synthetic landscape and of the natural example in the Eastern Alps show a highly bimodal landscape with the highest channel slopes at steep valley flanks and tributary headwalls, whereas areas above maintain rather low channel slopes. Thus, the inflection point in the channel slope–elevation distribution around the ELA is rather due to reducing the proportional fraction of topography below and above than a net accumulation of flat surfaces around.

Mountain ranges worldwide (e.g. European Alps, Andes, Tien Shan, Pyrenees) are characterized by a similar transition from increasing to decreasing channel slopes with elevation, while channel slopes increase continuously in others (e.g. Atlas, Carpathians) (for review Robl et al., 2017). In addition to the glacial steepening of low-lying landscape parts modelled and observed in our study (Figures 4 and 9), other influences have been suggested to cause such contrasting channel slope distributions (e.g. Hergarten et al., 2010; Künni & Pfiffner, 2001; Robl et al., 2015a).

According to the hypothesis of fluvial prematurity, low-gradient high-altitude surfaces are remnants of low-relief topography lifted by a recent uplift event (e.g. Hack, 1975; Hergarten et al., 2010). In this case, gentle channel slopes and valleys form the summit domains, while the regions below are steep and incised, as the two contrasting landscape types represent the ancient and recent tectonic regime, respectively. This type of elevated low-relief surfaces is not related to the glacial extent, shows a lower interquartile range than glacial landforms below and above, as the two contrasting channel slope distributions (e.g. Hergarten et al., 2010; Kühni & Pfiffner, 2001; Robl et al., 2015a).

5.3 Channel slope–elevation distributions of the Eastern Alps

The channel slope–elevation characteristics should in principle allow the interpretation of the degree of glacial imprint of entire mountain ranges. During the LGM, the Eastern Alps featured a continuous transition from a fully glaciated western to a fully unglaciated eastern part. The varying degree of glacial imprint is deducible today by analysing the channel slope–elevation distributions of individual catchments. The three catchments analysed in this study, representing the minorly glaciated, partly glaciated and fully glaciated topography of the Eastern Alps, show the temporal transition from fluvial to glacial characteristics (which progressively evolve in models (Figure 4a–c and 9) in spatial vicinity in the real world (Figure 8f–h). While the fluvial Trofaiach catchment shows a monotonic increase in mean channel slope with elevation (Figure 8f), a transition from increasing to decreasing channel slope becomes increasingly distinctive from the partly to fully glaciated catchment (Figure 8g, h).

However, a similar pattern but with a significantly lower interquartile channel slope range also emerges at the eastern fringe of the Alps outside the LGM ice extent (Robl et al., 2015a). This advocates the interpretation of a premature landscape induced by a tectonic driver (Hergarten et al., 2010; Legrain et al., 2014; Robl et al., 2015a; Wagner et al., 2010). In this part of the Eastern Alps, the vertical position of the turning point in the channel slope–elevation distribution varies. In contrast, the glaciated realm of the Eastern Alps shows significantly higher interquartile channel slope ranges and the turning point, generally located at about 1800 m (Figure 8h), roughly marks the altitude of the ELA (Robl et al., 2015a). This highlights the diagnostic capabilities of channel slope–elevation distributions.

5.4 Limitations

To study the feedbacks between glacial dynamics and topographical development, we kept our experiments as simple as possible and neglect several processes that influence landscape evolution. For example, fluvial erosion where glaciers are absent (e.g. Schlunegger & Norton, 2013), tectonic forcing (e.g. Sternai et al., 2013; Tomkin et al., 2003), stabilizing feedbacks in modelling erosion for adverse slopes within overdeepenings (e.g. Alley et al., 2003; Haeberli et al., 2016; Patton et al., 2016; Werder, 2016) and subglacial bedrock fracturing in response to high differential stresses (e.g. Leith et al., 2014) are not included.

Further, the erosion laws’ empirically determined proportionality constants and velocity exponents are particularly uncertain, ranging roughly from 0.6 (Cook et al., 2020) to 2 (e.g. Herman et al., 2015; Koppes et al., 2015). However, variations in parameters such as erosion law scaling constants only control the pace of erosion, while the spatial patterns of erosion are qualitatively similar (Braedstrup et al., 2016; Egholm et al., 2011, 2012a). Therefore, we focus our analysis
on the spatial and temporal patterns rather than absolute rates of erosion.

Hillslope diffusion is only included to maintain maximum bed channel slopes of 45°, which is considered the upper limit for iSOSIA’s validity (Egholm et al., 2011). This has a direct effect on the landscapes’ geometry. With continuously ongoing glacial erosion on the valley floor and persisting topography above, surface gradients are rising and the valley flanks tend to steepen beyond the slope threshold. As a consequence, hillslope processes are turned on to prevent the landscape from becoming too steep, and hence widen the valley with uniform flank side gradients of 45°. This results in channel slopes clustering around 45° within all altitude ranges (Figure 4) and a slight conversion of emerging U-shaped valleys towards more V-shaped valley cross-sections (Figures 3 and 6). However, the final modelled landscape (Figure 2b) possesses only a small surface fraction with bed channel slopes of 45°, and allowing the topography to rise above this threshold channel slope would not condemn the diagnostic capabilities of channel slope–elevation distributions and may even clarify the distinction between highly and less glacially modified landscape patches.

6 | CONCLUSIONS

Numerical modelling of glacial and periglacial erosion processes on a synthetic and a natural fluvially conditioned landscape combined with morphological analysis of topographic properties of the Eastern Alps leads us to the following conclusions:

The model results show that glacial erosion is focused in valleys of the fluvially conditioned drainage system, with the highest total erosion at trunk valleys (erosion hot spots). Peaks and ridges towering above the glacier network are only affected by periglacial erosion, which amounts to only a few metres (erosion cold spots). Thus, we observe a general increase in relief due to glaciation in these regions.

Spatial variations in erosion rate and the timing of maximum rates are dominated by valley overdeepening. Overdeepening first occurs in down-glacier reaches and progressively extends upstream. Controlled by the hydrological configuration of the glacier base, we observe a strong positive feedback between the formation of overdeepenings and the sliding rate. This makes valley excavation by glaciers a non-linear process.

In our simulation, glacial erosion leads to a highly bimodal landscape with the highest channel slopes at steep valley flanks and tributary headwalls, whereas areas above maintain their fluvial characteristics with rather low channel slopes gradients. Two transitions from increasing to decreasing channel slopes with elevation are emerging, one below (where overdeepenings dominate the channel slope distribution) and one above the ELA. In natural landscapes, the first turning point is rarely observable as overdeepenings are filled by water and sediment.

Glacial processes alter the (large-scale) topographic pattern of initial fluvially conditioned landscapes. Distributions of channel slope with elevation vary with the degree of glacial imprint, as demonstrated by the progressive evolution from a fluvial to glacial landscape in our numerical experiments. This temporal transition can also be observed in spatial vicinity from a fully glaciated western to a minorly glaciated eastern part of the Eastern Alps. Thus, channel slope–elevation plots serve as a diagnostic tool for interpreting the glacial–fluvial influence in mountain landscapes.

The position of the turning point and the variance of channel slopes in the channel slope–elevation plots serve to distinguish between drivers of the formation of bimodal landscapes. In a glacial landscape, the turning point relates to the ELA and shows a high interquartile channel slope range. On the contrary, in bimodal landscapes without glacial imprint, the vertical position of the turning point varies and the interquartile range is comparatively small.

While the topographic evolution of fluvially conditioned mountain ranges has been successfully explored by the application of stream power incision models (SPIMs) (e.g. Robl et al., 2017, and references therein), a high-resolution, numerical description of glacial mountain ranges and their evolution towards a hypothetical glacial steady state is still in its infancy. In contrast to the SPIMs, where fluvial erosion is computed from geometrical properties of the drainage system (i.e. local channel slope and upstream contributing drainage area), state-of-the-art glacial erosion models are computationally expensive as they describe erosion based on the physical process of moving ice over its bedrock (e.g. Braun et al., 1999; Egholm et al., 2011; Herman et al., 2011). These models face some limitations and stability issues (e.g. steep slopes) when considering the mathematical and numerical complexity of ice dynamics. While these models are extremely useful for describing landscape modifications in periods of fluctuating climate (e.g. during a glacial–interglacial cycle), computational performance and numerical stability impede high-resolution, large-scale simulations over the million years’ time scale. Recently and similar to the SPIM, a new approach to understanding glacial landscape evolution has been published (Deal & Prasicek, 2021), which allows the computation of glacial erosion rates from topographic properties and simple climate assumptions. Simultaneously, Hergarten (2021) presented a new landscape evolution model, describing glacial landscape evolution based on a stream power law for glacial erosion, and coupled this model with a fluvial erosion model taking into account both incision and sediment transport. Further benchmarking will show whether these models are able to produce large-scale landscape patterns observed in natural glacially shaped landscapes and readily described by higher-order ice flow models for small domains. However, if these rather simple models are similarly successful as their fluvial counterparts, we will apply this type of models and build on the findings of this study to explore the processes and topographic signatures during the transition from fluvial to glacial landscapes (and vice versa) for large mountain ranges with high spatial resolution over the million years’ time scale.

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COMPETING INTERESTS

The authors declare that they have no conflicts of interest.

AUTHOR CONTRIBUTION

M.L. configured the experiments, and prepared the manuscript with contributions from J.R. and G.P. D.L.E. provided the model iSOSIA and the supervision to use it. D.L.E., K.S., G.G. and S.H. contributed to the final version of the original draft. Acquisition of funding was done by J.R. and K.S. All authors have read and agreed to the published version of the manuscript.
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Additional supporting information may be found online in the Supporting Information section at the end of this article.

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