Modelling alpine glacier geometry and subglacial erosion patterns in response to contrasting climatic forcing

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Abstract
Climate exerts a primary control on glacier mass balance, driving changes in ice flux, which affects basal sliding and subglacial erosion. Although past glacier reconstructions have been widely used as paleo-climate proxies, extracting quantitative temperature/precipitation data from paleo-glaciers is still challenging. The iSOSIA ice model was used over a synthetic landscape to quantitatively investigate the non-linear dependence of glacier geometry and ice dynamics on climate forcing, as well as to analyse how spatial patterns of subglacial erosion and erosion rates vary between the accumulation and ablation zones for different climatic settings. We performed 10 calibrated climatic scenarios with different combinations of temperature and precipitation rate in two separate sets of simulations [maintaining either same equilibrium line altitude (ELA) or same ice extent; SA and ST, respectively]. Our results reinforce the role of climate and the importance of ice flux in conditioning glacier geometry. In SA simulations, contrasting climatic scenarios showed a major influence on glacier length and thickness. In ST simulations, calibrated ELAs presented a variability of over 300 m in order to maintain similar ice extent, but showed reduced changes in glacier thickness. These results highlight the importance of ice flux to predict glacier geometry, when compared to the overall mass balance from an ELA estimate. Subglacial abrasion and quarrying showed notable differences between the accumulation and ablation zones for varying climatic forcing, with multimodal erosion distributions for increased precipitation/temperature, shifting erosion towards higher-elevation tributaries and connecting erosion patches in the ablation zone. Such trends highlight the role of subglacial hydrology and ice-bed contact (cavitation) in dictating patterns of subglacial erosion, while the depth of erosion relates closely to ice flux and climate inputs. Our results have potential implications for paleo-climate reconstructions based on glacier geometry and provide insights on how subglacial erosion can help estimate paleo-climatic conditions from paleo-glacial records.

KEYWORDS
(paleo-)ELA, climatic forcing, glacier geometry, ice dynamics, numerical modelling, subglacial erosion patterns

1 | INTRODUCTION

Having shaped polar and alpine landscapes over the Neogene and Quaternary, glaciers are known to be major geomorphic agents acting on the Earth’s surface (e.g. Cook et al., 2020; Hallet, 1979; Herman et al., 2015, 2021; Sugden et al., 2017). Glacier dynamics and paleo-geographical reconstructions have also been used for estimating past changes in climate, but limitations remain for inferring past surface temperatures and precipitation patterns (e.g. Dahl & Nesje, 1992; Kerschner & Ivy-Ochs, 2008; Pearce et al., 2017; Protin et al., 2019).

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Paleo-climatic inference from past glacier reconstructions (i.e. ice extent and thickness from geomorphological markers such as moraines or trimlines) can be difficult to convert into past changes in temperature or precipitation (e.g. Kerschner & Ivy-Ochs, 2008), due to the multiple controlling factors including local topography and climatic conditions (e.g. Kessler et al., 2006; Norton & Hampel, 2010), long-term climatic trends (Oerlemans, 1989, 2005) as well as internal climate variability (Roe, 2011; Roe & O’Neal, 2009). These factors can affect local glacier mass balance and potential trade-offs between climatic parameters (e.g. Allen et al., 2008; Becker et al., 2016; Blard et al., 2007; Protin et al., 2019; Solomina et al., 2016). For some specific settings, this limitation can be overcome by using other independent local paleo-climactic proxies such as paleo-lake levels to constrain past precipitation and temperature changes (Martin et al., 2020; Mey et al., 2020). Thus, despite the application of paleo-climate reconstructions and especially paleo-ELAs (equilibrium line altitudes) based on paleo-glacial landforms, there have been only few attempts to estimate the influence of precipitation changes on past glacier geometry or inferred paleo-ELAs (e.g. Kessler et al., 2006; Lai & Anders, 2021; Reixach et al., 2021).

Precipitation and temperature are primary climatic controls on glacier mass balance, driving basal sliding and subglacial erosion and changes in both ice thickness and extent (e.g. Hallet, 1979; Herman et al., 2015, 2021; Koppes et al., 2015). However, still little is known on how subglacial erosion patterns (spatial trends and differences) behave under temporally or spatially varying combinations of temperature/precipitation, even for fixed glacier geometry or ELA configurations. Indeed, such climatic complexity may be reinforced by non-linear effects of hydrology on ice dynamics and resulting rates and patterns of erosion. Although erosion is generally assumed to scale with ice sliding velocity based on Hallet’s abrasion law (Hallet, 1979), abundant meltwater is capable of speeding up sliding and therefore subglacial erosion (e.g. Bernard et al., 2020; Egelholm et al., 2012; Herman et al., 2011; Ugelvig et al., 2016, 2018). In addition, the ice-bed contact friction plays a crucial role in controlling subglacial erosion (e.g. Schoof, 2005), which in turn depends on the water flux, ice thickness and efficiency of upglacier quarrying. Therefore, ice dynamics and resulting subglacial erosion patterns may also be indicative of a specific climatic setting, particularly a minimum level of precipitation (Koppes et al., 2015).

Accessing subglacial environments and obtaining direct onsite measurements of subglacial processes and erosion rates are challenging, and surface process models (SPMs) have therefore been used for decades to simulate ice dynamics and subglacial processes (Harbor et al., 1988; Oerlemans, 1984; Tomkin & Braun, 2002), demonstrating their usefulness for investigating the role of surface processes in long-term landscape evolution. Nevertheless, the complexity imposed by multiple processes and possible feedbacks between glacier dynamics, climate and topography/tectonics (Anderson et al., 2006) have been tackled only recently with more sophisticated and numerically efficient modelling approaches (e.g. Braun et al., 1999; Egelholm et al., 2009; Herman & Braun, 2008; Kessler et al., 2006; Tomkin, 2009). Ice-model developments from 1D simulations (e.g. MacGregor et al., 2000; Oerlemans, 1984) to more recent 2D and 3D higher-order models (Egelholm et al., 2011; Jouvet et al., 2008; PISM—Winkelmann et al., 2011; ISSM—Lair et al., 2012; Elmer ice—Gagliardini et al., 2013) allow investigating large-scale, long-term glacier dynamics by also incorporating the effects of rugged topography. More recently, higher-order ice-flow models coupled with bedrock erosion, subglacial hydrology, sediment transport and deposition (Bernard et al., 2020; Egelholm et al., 2012; Ugelvig & Egelholm, 2018; Ugelvig et al., 2016) numerically reproduced topographic widening, hillslope processes and more complex hydrological feedbacks. By simulating landscape evolution over thousands to millions of years, these models allow us to run virtual experiments and explore how changes in climate, tectonics and topography interact, which is otherwise highly challenging to investigate in a laboratory or in nature.

The existing links between paleo-ELAs and (paleo-)glacier extent have also been investigated using ice models (e.g. Blard et al., 2007; Protin et al., 2019). However, for fixed ice extent or glacier ELA, different climatic scenarios (i.e. combinations of precipitation/temperature) can result in significant variations in either inferred ELA (e.g. up to ~150 m for Younger Dryas alpine glacier; Protin et al., 2019) or ice geometry, especially between the accumulation and ablation zones, challenging a direct inference of paleo-climates. Moreover, such differences can also highly affect ice dynamics (ice flux, subglacial hydrology) and, consequently, be observed on subglacial erosion patterns (e.g. Lai & Anders, 2021).

To investigate this question, we used the three-dimensional Integrated Second-Order Shallow Ice Approximation model (iSOSIA; Egelholm et al., 2011) to numerically explore how both ice dynamics and erosion patterns are influenced by specific climatic scenarios (i.e. precipitation and temperature conditions). Using a kilometre-scale synthetic alpine landscape, we performed two sets of simulations with a selected range of climate scenarios preserving either similar glacier ELAs or ice extents (10 scenarios each). By integrating in our simulations abrasion and quarrying laws as subglacial erosion, as well as the role of subglacial hydrology in cavities or channels, we investigated how ice dynamics (i.e. ice flux; subglacial velocities and ice thickness) and erosion rates vary spatially and differ between the accumulation and ablation zones under contrasting climatic conditions. We also analysed how glacial erosion patterns and their relative magnitudes evolve in each scenario, based on a reference scenario and a set of forcing parameters. With this approach, our goal is eventually to discuss the potential and limitations in using paleo-glacial reconstructions as a paleo-climatic proxy, but also ultimately how subglacial erosion patterns can be used as complementary proxies for paleo-climate studies.

2 | METHODS AND MODEL SETUP

2.1 | Ice model iSOSIA

In the present study, we use the ice model iSOSIA, a finite-volume solver with explicit time integration that allows the computation of ice flow, subglacial hydrology and sediment transport (Egelholm et al., 2011, 2012). The numerical integration of ‘higher-order’ effects ensures the preservation of good accuracy compared to full Stokes’ model predictions for steep and rugged topography (Egelholm et al., 2011), while allowing long-term and large-scale simulations necessary to investigate paleo-glacial questions and the impact of ice dynamics on erosion and mountain relief (Egelholm et al., 2012).
In iSOSIA, the climatic input is based on a simple mass-balance approach using a positive-degree-day (PDD) model (here we use a snow/ice melting degree factor of $3 \times 10^{-3}$ m w.e. °C$^{-1}$ day$^{-1}$; see Table 1), which is a function of temperature (decreasing linearly with altitude at a lapse rate of $6$ °C/km, with an annual temperature amplitude $\Delta T_a$ of $10$ °C (Pedersen et al., 2014)) and precipitation. We consider a single melting degree factor for both ice and snow, since all precipitation is turned into ice after accumulation (based on yearly average rates). In our simulations, precipitation is considered spatially uniform. Any change in ice thickness is therefore computed as a balance between ice flow, ice ablation and accumulation [Equation 1], while mass conservation assumes a constant ice density spatially and at depth. Snow avalanching in iSOSIA (Scherler & Egholm, 2020) transports new accumulated snow downward towards lower elevation cells, as long as ice surface slope between the cells is larger than $\alpha_{\text{aslope}}$ (fixed to 0.5 m/m in our simulation, see Table 1).

Ice fluxes are computed in iSOSIA by vertically integrating horizontal ice-flow velocities at cell boundaries. In addition, bed topography, ice thickness and elevation are averaged in each grid cell (Braedstrup et al., 2014). Ice thickness changes are derived from the following mass conservation law:

$$\frac{\partial h_{\text{ice}}}{\partial t} = -\nabla \cdot \bar{F} + M$$

(1)

where $h_{\text{ice}}$ is the thickness of ice (m), $t$ is time (years), $\bar{F}$ is the horizontal flux (the product of ice thickness and the depth-averaged ice velocity vector, $m^2$ year$^{-1}$) and $M$ the mass source term (m/year). This equation is integrated for each model cell using a flux limiter that prevents negative ice thicknesses.

Horizontal stress components ($s_{xx}$, $s_{yy}$, $s_{xy}$) are based on Glen’s flow law (stress exponent $n = 3$) and horizontal flow velocities depend non-linearly on ice-surface gradients and curvature (Egholm et al., 2011). In iSOSIA, the horizontal stress components are assumed not to vary at depth within the ice, and constant values can be computed from depth-averaged strain rate components. In that sense, depth-averaged velocities depend on the local ice thickness, ice surface gradient, local variations in depth-averaged ice velocities and material parameters (note that ice is assumed to be isotropic in iSOSIA).

In our simulations, we used a modified sliding equation (see Table 1 for parameter details), based on Ugelvig et al. (2018):

$$U_b = \frac{C_s + \tau_b^3}{(1 - SF)^{1/2}}$$

(2)

where $U_b$ is the sliding velocity (m/year), $C_s$ is the sliding coefficient (Pa s$^{1/2}$), $\tau_b$ is the computed basal shear stress (Pa; Egholm et al., 2011) and $SF$ the ice-bed contact fraction that accounts for the cavities ($SF = S/L_v$, $S$ being the cavity length and $L_v$ the cavity spacing; Ugelvig et al., 2018).

We follow an approach of subglacial hydrology modelling described by a system of subglacial channels or connected cavities, and assume that meltwater is transported at each grid cell instead of each individual channel or cavity. Therefore, mean cavity and channel spacing are set as constants (Table 1), while the height of the topographic step ($h_s$) is modelled based on slope and ice-flow direction (e.g. Ugelvig et al., 2016, 2018). Water is generated by ice melting at the surface and at the ice bed. Water flow follows the same direction as the ice flow (depth-averaged) at any point in the glacier. In our approach we do not account for seasonal or diurnal pressure variations (as in Ugelvig et al., 2018), but rather we model a steady-state solution, where steady-state effective pressures (difference between ice and water pressures at the bed in cavities or channels) are computed and the highest effective pressure (corresponding to

| Mass balance | Lidate | Temperature lapse rate with elevation | $6 \times 10^{-3}$ °C m$^{-1}$ |
|--------------|--------|-------------------------------------|-------------------------------|
| mPDD         | Melting degree-day factor            | $3 \times 10^{-3}$ m w.e. °C$^{-1}$ day$^{-1}$ |
| $\Delta T_a$| Annual temperature amplitude        | 10 °C (Pedersen et al., 2014) |
| $C_s$        | Sliding coefficient                  | $5 \times 10^{-3}$ Pa s$^{1/3}$ |

| Erosion | Abrasion coefficient | $4 \times 10^{-7}$ m$^{-1}$ year$^{-1}$ |
|---------|----------------------|----------------------------------------|
| $K_a$   | Quarrying coefficient | $2 \times 10^{-6}$ m$^{-1}$ year$^{-1}$ |

| Hydrology | Minimum hydraulic conductivity | 0.01 kg$^{-1/2}$ m$^{3/2}$ |
|-----------|--------------------------------|----------------------------|
| $L_v$     | Latent heat of fusion          | $3.4 \times 10^4$ J kg$^{-1}$ |
| $L_v$     | Cavity spacing                 | 4 m (Bernard et al., 2020) |
| $L_v$     | Channel spacing                | 200 m (Bernard et al., 2020) |
| $\alpha_{\text{aslope}}$ | Critical accumulation slope | 0.5 (m/m) |
| $A$       | Ice flow parameter            | $10^{-16}$ Pa$^{-3}$ year$^{-1}$ (Egholm et al., 2012) |
| $\rho_{\text{ice}}$ | Ice density                | 910 kg m$^{-3}$ |
| $g$       | Acceleration due to gravity   | 10 N kg$^{-1}$ |
| $n$       | Ice creep constant            | 3 |
| $B$       | Ice viscosity constant        | $7.33 \times 10^3$ Pa s$^{1/3}$ |
| $\beta$  | Cavity shape parameter        | 0.7 (Anderson, 2014) |
the lowest water pressure) is taken, classifying the grid cell as cavity or channel dominated. In cavities, the effective pressure, \( N_{\text{cav}} \), is calculated as

\[
N_{\text{cav}} = \frac{B}{C^{3}} \pi^{1/8} \left( \frac{h_{s}}{C^{18} C^{19}} \right) \quad \text{where } B \text{ is the ice viscosity constant (Pa s}^{1/3}, \text{)} \quad n \text{ is the ice creep constant (} n = 3) \quad \text{Ub is the sliding velocity (m/yr)}, \quad h_{s} \text{ is the height of the topographic step (m) and } S \text{ the cavity length (m) (see Table 1 for parameter details and Figure 1 for an illustration).}
\]

In channels, the effective pressure is

\[
N_{\text{cha}} = nB \left( \frac{L_{c} \times q_{w}}{\rho L_{s} A_{C}} \right)^{0.8} \quad \text{(4)}
\]

where \( L_{c} \) is the mean channel spacing (m), \( q_{w} \) is the water flux under the glacier (m\(^2\) s\(^{-1}\); Ugelvig et al., 2018), \( Q_{w} = \rho g h_{ic} \) is the gradient of the hydrological head (Pa m\(^{-1}\)), \( \rho \) and \( L_{c} \) are ice density and latent heat and \( A_{C} \) the channel cross-section (m\(^2\)) [see Equation 3 for parameter description and Table 1 for details]. The sizes of cavities and channels are modelled as

\[
S = \frac{1}{\beta} h_{s} \left( \frac{L_{c} \times q_{w}}{k_{\text{min}} \times \sqrt{Q_{w}}} \right)^{0.8} \quad \text{and}
\]

\[
A_{C} = \left( \frac{L_{c} \times q_{w}}{k_{\text{min}} \times \sqrt{Q_{w}}} \right)^{0.8} \quad \text{(6)}
\]

where \( \beta \) is the constant cavity shape parameter, \( k_{\text{min}} \) is the minimum hydraulic conductivity (kg s\(^{-1/2}\) m\(^{3/2}\)) and \( L_{s} \) the mean cavity spacing (m). The latter is fixed in this work, which uses a steady-state solution for subglacial hydrology (following Bernard et al., 2020). \( L_{c}^{*} q_{w} \) and \( L_{c}^{*} q_{w} \) are water discharges, respectively, in a cavity and a channel. For previously presented parameters, see the description for Equation 3 and the details in Table 1.

From Equations 3–5, cavity size controls the effective pressure (\( N_{\text{cav}} \)) of the system influencing, in turn, basal sliding speed [Equation 2].

In our simulations, subglacial erosion (MacGregor et al., 2009) is computed as the sum of two components: abrasion and quarrying. Abrasion accounts for the direct effect of ice-to-bedrock friction and adopts Hallet’s abrasion law (Hallet, 1979), in which abrasion rates scale with sliding velocity squared [Equation 7]. The adopted law assumes topographic steps have roughness, since the effective ice-bed contact ratio (\( SLf \)) is considered within the sliding velocity [Equation 2]. In the simulations, we did not consider the potential shielding effect of basal sediments nor the tool effect of basal debris (Ugelvig & Egholm, 2018), and the computation of abrasion follows...
where $K_a$ is the abrasion coefficient ($m^{-1}$ year) and $U_b$ is the basal sliding velocity (m/year). $K_a$ depends on basal-ice debris concentration, debris-particle shape and the relative hardness of debris particles and bedrock (Hallet, 1979). We do not model variations in this property and treat $K_a$ as a constant (see Table 1 for details).

Quarrying accounts for plucking of bedrock pieces and is influenced by first-order controls such as sliding velocity (which also considers ice-bed contact fraction; Equation 2), effective pressure and the presence of pre-existing fractures (Iverson, 2012). Quarrying depends on the possibility of opening a pre-existing fracture in a rock that is in contact with the ice but nearby a cavity without ice-bed contact (e.g. Bernard et al., 2020). In general, this assumes that the probability of quarrying a topographic step next to a cavity increases with the ice-bed contact area of the step. We computed quarrying based on Iverson (2012) and follow the approach of Ugelvig et al. (2016), which also investigated the controlling effect of bed slope in quarrying:

$$E_q = K_q \times N^3 \times U_b \times (\nabla b_i)^2$$

where $K_q$ is the quarrying coefficient, representing the effect of bedrock lithology and pre-existing fractures and used for scaling the erosion rates. $N$ is effective pressure, which is the highest of $N_{air}$ or $N_{inv}$ (i.e. the hydrological system with lowest water pressure dominates the hydrological system; see Equations 3 and 4). $\nabla b_i$ is the bed slope in the direction of sliding (see Table 1 for parameter details).

2.2 Modelling approach and setup

In this study, our overall aim is to numerically investigate the non-linear effects of climatic forcing on (1) glacier geometry (ice extent and thickness, area and volume), (2) ice dynamics (basal sliding velocity and subglacial hydrology) and finally (3) subglacial erosion by abrasion and quarrying (spatial patterns and magnitudes).

Given the non-linear dependence of ice geometry and dynamics on surface mass balance, we designed 10 climatic scenarios based on different combinations of annual precipitation rate and sea-level temperature as input climatic parameters. Furthermore, we tested two different modelling approaches in order to put our study into a broader paleo-glaciological context: (1) simulations with fixed ELAs and varying simulated ice geometries (same-ELA simulations, hereafter referred to as ‘SA’) and (2) simulations with fixed ice extent (i.e. similar ice front position) but varying simulated ELAs (same-ice-extent simulations, hereafter referred to as ‘ST’). Therefore, the input sea-level temperature is varied for a fixed range of precipitation in order to maintain either the ice terminus or the ELA, which are fixed in ST or SA simulations, respectively (Figures 2c and d). These two sets of simulations reflect different approaches in paleo-glaciology, where paleo-ice reconstructions are based either on geomorphological markers to infer past ELAs or simulated from ELA estimates from other independent paleo-climatic proxies (e.g. Dahl & Nesje, 1992; Kerschner & Ivy-Ochs, 2008; Pearce et al., 2017; Protin et al., 2019). We ran iSOSIA simulations over a synthetic fluvial landscape, with $20 \times 40$ km total extent and 100 m cell resolution. The synthetic landscape has an elevation range between 0 and 4500 m (Figures 3a and b), similar to a typical alpine landscape. The model setup is performed as described above, and the main setup parameters for SA and ST simulations are described in Table 1.

First, we designed a ‘reference’ climatic scenario, selecting precipitation/sea-level temperature input (1 m/year and 5.5°C, respectively), which translates into a given ELA (1985 m; Figure 1a) and ice extent (30 km long; Figures 3b and c). This ELA value is roughly at the hypsometric mode of the synthetic landscape (Figure 3a), and moreover it falls into the range of reported alpine paleo-ELAs for the last glacial cycle (e.g. Kuhlemann et al., 2008; Višnjević et al., 2020). Preliminary sensitivity tests showed that the behaviour of temperature vs. precipitation in order to maintain the same mass balance does not significantly change for different ELA values; therefore, we consider that our ELA choice does not limit or alter our modelling investigation regarding the ice-dynamics response to climatic forcing. Annual precipitation scenarios were imposed and range within possible present-day and alpine precipitation estimates for the last glacial cycle (Allen et al., 2008; Becker et al., 2016; Isotta et al., 2014; Višnjević et al., 2020), including end-member scenarios of 0.25 and 4 m/year. For each precipitation scenario, the annual sea-level temperature was constrained by a trial-and-error process (Figure 2c) to predict either similar ELAs at 1985 m (SA simulations; Figure 2a) or similar ice extents of 30 km long (ST simulations; Figure 2b). For ST simulations, we imposed a maximum 100 m difference (i.e. one model cell) in ice-front position for fitting the different scenarios (Figure 3c).

For each calibrated climatic scenario and associated surface mass balance (Figures 2a and b), ice simulations were run for 1.7 kyr without subglacial erosion (i.e. abrasion and quarrying off), to ensure that a steady-state configuration was attained with stable ice geometry and only small ice-thickness changes, mostly coming from snow avalanche and subglacial hydrology (small variations in effective pressure). The steady-state ice configurations were checked with ice-thickness changes below 0.3% over the last 500 years of the total simulation run. Subsequently, steady-state ice configurations were used as initial input to run simulations over another 1 kyr with subglacial erosion enabled (i.e. abrasion and quarrying on), and where $K_a$ and $K_q$ (Table 1) were calibrated for the reference scenario (1 m/year precipitation) in order to provide maximum abrasion and quarrying rates of 1 mm/year each (reference rates for comparison with subsequent simulations).

Output ice configurations (glacier thickness and extent), predictions regarding ice dynamics (basal sliding and deformation velocities, subglacial hydrology) and erosion (abrasion and quarrying spatial trends) were extracted for further investigation. To ensure quantitative comparisons between the different climatic scenarios, we only kept from simulation outputs the trunk glacier and its connected tributaries (Figure 3a), while removing non-connected tributaries for subsequent analysis. For this step, we used an automated Matlab® function (bwrapper) to track only the connected cells based on neighbours search, imposing a 25 m threshold for ice thickness to also remove small areas with potential ice-mass redistribution instabilities (mostly located on valley flanks or at crests). Given the large variability in the input temperature/precipitation scenarios (Figure 2c), the connectivity between the trunk glacier and some tributaries varies between the different climatic scenarios.
(Figure 3a), with some tributaries joining when increasing (decreasing) annual precipitation for SA (respectively ST) simulations (see the next section for details and consequences on ice geometry). Such variability in tributary connectivity creates a more complex but realistic ice simulation, as also suggested by Bernard et al. (2020).

3 | SIMULATION RESULTS

3.1 | Climatic scenarios and ice-extent/ELA predictions

The mass-balance calibrations for the two simulations (SA and ST) and 10 climatic scenarios (input annual precipitation between 0.25 and 4.00 m/year, with a reference scenario of 1 m/year; Figures 2a and b) resulted in a range of annual sea-level temperatures (Figure 2c) that compensate the varying precipitation and maintain the same ELA (SA, 2.0–10.2°C) or the same ice extent (ST, 3.5–9.6°C). In addition, we observed similar temperature/precipitation relationships between SA and ST simulations, despite the different simulation approach (Figure 2c). Our calibrations highlighted the impact of drier/colder or more humid/warmer conditions on both the ELA and ice-extent variability (Figure 2d). The different climatic scenarios resulted in ice-extent ranges of 15–30 km (SA simulations) and ELAs of 1750–2090 m (ST simulations). Moreover, our results highlight that ice simulations are more sensitive to climate drying/cooling compared to the reference scenario (1 m/year).

In the following, we compare the 10 calibrated climatic scenarios from each of the two simulations (SA and ST) with the overall aim to evaluate the climatic impacts on both glacier geometry and erosion patterns over a synthetic landscape (Figure 3a). We present our results with comparisons between the accumulation (above the ELA) and ablation (below the ELA) zones separately, as they show significant differences and provide further insights into glacial dynamics. We also indicate the reference scenario with 1 m/year of precipitation and 5.5°C for comparison purposes between colder/drier or hotter/wetter climatic scenarios (Figure 2) and refer to the scenarios by their precipitation for a more concise text, although the precipitation/temperature pair should always be considered (see Figures 2c and d for reference).

3.2 | Outputs for ice geometry

Same-ELA (SA) simulations show significant differences between climatic scenarios, with up to 15 km changes in glacier length. This

![Figure 2](https://wileyonlinelibrary.com)
change in glacier geometry corresponds to a relative change of \( \sim 60\% \) in relation to our reference scenario, and highlights the importance of integrating ice flux rather than just the mass balance for predicting ice geometry (Figures 3a and b). The incorporation of newly connecting tributaries to the trunk glacial system also reinforces the observed geometric differences, and adds a more realistic approach to a seldom analysed aspect. For SA simulations, ice thickness also shows the biggest change for our simulations, with differences of up to 700 m between the two end-member climatic scenarios (Figure 3b). In contrast, same ice-extent (ST) simulations present much smaller differences (up to 100–150 m difference; Figure 3c), even if the calibrated ELAs show a large variability between 1750 and 2100 m for the different climatic scenarios (Figure 3c).

Glacier cross-sections in the ablation (Figure 4a) and accumulation (Figure 4b) zones confirm the ice-thickness changes between scenarios, but also reveal changes in the ice-surface profiles for the two areas. In the ablation zone (cross-section 1 in Figure 4a), ice-thickness changes are roughly constant along the transverse profile and the ice surface is apparently convex in shape. In contrast, glacier cross-sections in the accumulation zone show significant variations in ice-surface changes along the transverse profile, with maximum differences in the central part of the trunk glacial valley, but almost none in the valley margins (i.e. at the bedrock ice contact; Figure 4b).

Our results also reveal interesting behaviours for tributary dynamics. For SA simulations, increased precipitation scenarios promote glacier advance, both for the trunk glacier and its tributaries that progressively join the trunk glacial system (Figure 3a). The opposite behaviour is observed for ST simulations, with disconnected tributaries for increased precipitation scenarios. Such climatic scenarios are associated with increased temperatures in order to

**FIGURE 3** (a) Model extent (40 \( \times \) 20 km, 100 m resolution) and synthetic fluvial landscape. Inset shows hypsometric distribution (black dashed line is reference ELA at 1985 m). The main glacier system is represented for the reference (1 m/year) and end-member (0.25 and 4 m/year) climatic scenarios (SA simulations, showing large variations in ice-front position between scenarios). Note the star symbols (a and b) that locate the tributary junctions for SA (a, b) and ST (b) simulations. See main text and Figures 5 and 6 for details. Longitudinal (A–A’ in Figures 3b and c) and cross-section (1 and 2 in Figure 4) profiles are also indicated. (b, c) Longitudinal ice profiles and ELA positions along the trunk valley (A–A’ in Figure 3a) for SA (b) and ST (c) simulations considering different climatic scenarios (coloured lines) [Color figure can be viewed at wileyonlinelibrary.com]
maintain the same ice extent for the trunk glacier, and cause disconnection of tributaries for which the integrated mass balance is not maintained.

Spatial changes in glacier area are noticeably different between the accumulation and ablation zones, with a higher variability in the ablation zone for both SA and ST simulations (Figures 5a and b). Given our approach for SA simulations (in which the ice extent can vary between scenarios), we observe glacier area changes by up to 100% between the two end-member scenarios. In ST simulations (in which the ice extent is fixed, but the ELA can vary between scenarios), variations in glacier area are smaller, but can still represent a 35% change due to changes in ice thickness, lateral glacier extent and tributary junction between end-member scenarios (Figure 5b). In the accumulation zones, glacier area changes are more limited, especially for ST simulations (<10% change; Figure 5b), but large ice-thickness variations and tributary junctions for SA simulations result in 50–60% change in glacier total area (Figure 5a).

We further investigated the glacier area change between accumulation and ablation zones by calculating the output AAR (accumulation area ratio, computed as the glacier’s accumulation area

![Figure 4](image1.png)  
**Figure 4** Ice-surface cross-sections in the ablation (a) and accumulation (b) zones (cross-sections 1 and 2, respectively in Figure 3a). Presented cross-sections are model outputs from ST simulations (coloured lines indicate the different climatic scenarios), although SA showed very similar ice configurations. Note the ice-surface convex (ablation zone) and concave (accumulation zone) shapes along the cross-sections. See text for details [Color figure can be viewed at wileyonlinelibrary.com]

![Figure 5](image2.png)  
**Figure 5** Main glacier system area changes (% compared to reference scenario with 1 m/year precipitation) for SA (a) and ST (b) simulations. Both accumulation (blue stars) and ablation (red circles) zones are represented, as well as thresholds in area change for the tributary junction (black arrows a-b; see Figure 3a for locations and text for details). The evolution of the output AAR is also shown (yellow crosses, right scale) [Color figure can be viewed at wileyonlinelibrary.com]
divided by its total area; Meier & Post, 1962) for each climatic scenario. Interestingly, the AAR changes by around 10–20% between climatic scenarios (Figure 5), both for SA and ST simulations, with higher values (AAR around 0.6) for dry/cold scenarios compared to wetter/warmer scenarios (AAR around 0.5).

Finally, we also checked the relationships between glacier area and volume, focusing on potential differences between accumulation and ablation zones for ST and SA simulations. Although these variables are not independent they may indicate different behaviours, illustrating also changes in overall glacier thickness. First, as also observed for glacier area (Figure 5), ice-volume differences are greater in the ablation zone for both simulations (Figure 6) and also higher for SA than ST simulations. For SA simulations (Figure 6a), we observe almost linear relationships between ice area and volume changes (both for accumulation and ablation zones) for dry scenarios (<1 m/year precipitation), showing a first-order control of ice area changes on volume variations. For wet scenarios (>1 m/year precipitation), this relationship is altered and ice volume increase less than area, with a clear difference between accumulation and ablation zones. This highlights the importance of ice-thickness changes in such
scenarios. For ST simulations, a very different behaviour between the accumulation and ablation zones is observed (Figure 6b). While there is an apparent linear relationship between area and volume changes for the ablation zone, for the accumulation both area and volume changes are limited (<10%), with no apparent relationship between the two metrics.

3.3 Outputs for subglacial erosion patterns

In this section, we present erosion results for ST simulations, since these have similar ice extent and allow for direct comparison of subglacial erosion patterns (spatial trends). We present erosion results for SA simulations in the online Supporting Information (Figures S1 to S6).

Mean abrasion rates are largely similar within the accumulation and ablation zones (Figure 7a), with an almost linear increase in mean abrasion for increasing precipitation. In contrast, there is a clear difference for quarrying outcomes, both in terms of their relationship with precipitation and their variability between accumulation/ablation zones: mean quarrying rates are higher in the ablation zone (two to three times) and quickly reach a plateau at around 1.5 m/year with increasing precipitation (Figure 7b). Abrasion and quarrying were calibrated for the reference scenario (1 m/year precipitation) to predict maximum rates of around 1 mm/year each; although they present similar mean values in the ablation zone (around 0.1 mm/year; Figure 7), mean quarrying is much lower in the accumulation zone (around 0.03 mm/year) and shows strong spatial variability in this zone.

We also computed the combined abrasion and quarrying rates to estimate the total eroded volume over 1 kyr for each climatic scenario. Although glacier extents are similar for ST simulations, we observe that the total eroded volume increases nearly 10-fold with precipitation (Figure 8a), which is a similar magnitude change to SA simulations (in which glacier area and volume are changing significantly; see Figures 6 and S2). Mean erosion rates are higher in the ablation zone (Figure 7), but the accumulation zone shows a slightly greater total eroded volume for wetter/warmer scenarios (Figure 8a), which can be explained by the relative increase in the area occupied by ice within the accumulation zone for ST simulations (lowering of the ELA for increased precipitation/temperature) in addition to the larger variability in quarrying rates for the accumulation zone (Figure 7b). Moreover, we investigated the relationship between ice volume and total eroded volume, which shows different behaviours between the accumulation and ablation zones: even if the ice volume is roughly constant in the accumulation zone, the total eroded volume still increases significantly, whereas it scales exponentially with ice volume in the ablation zone (Figure 8b).

After investigating large-scale integrated erosion rates and volumes at the scale of the entire glacier, we now explore the variability of erosion (abrasion and quarrying) at smaller scale. We computed distribution histograms of subglacial erosion (abrasion, quarrying and total erosion; Figure 9) for three end-member climatic scenarios: 1 m/year (reference), 0.25 m/year (lowest precipitation/temperature) and 4 m/year (highest precipitation/temperature). Our aim is to evaluate the evolution of local subglacial erosion from dry/cold to wet/warm scenarios. The output abrasion rates (Figures 9a–c) show similar distributions and magnitudes between the accumulation and ablation zones for the high-precipitation scenario (Figure 9c), while abrasion is clearly lower in the accumulation zone for the low-precipitation scenario (Figure 9a). In addition, we observe a clear trend towards higher abrasion rates with increased precipitation, as also evidenced for glacier-averaged abrasion rates (Figure 8a).

Quarrying rates show a similar trend between climatic scenarios, with a clear bimodal distribution in the accumulation zone for the low-precipitation scenario (Figure 9a). Furthermore, and different to the abrasion distributions, quarrying in the accumulation zone still presents much smaller values than in the ablation zone, even in the high-precipitation scenario (Figure 9f). This highlights an increased importance of quarrying in the ablation zone relative to the accumulation zone with increased precipitation.
The distribution of total erosion (Figures 9g-i) shows that with our model settings, abrasion provides, in general, the highest contribution to overall erosion, with very similar distributions between abrasion (Figures 9a-c) and total erosion (Figures 9g-i). This results from the large spatial variability in quarrying rates (Figures 9d-f) and our model setup, where we scaled maximum abrasion and quarrying to 1 mm/year for the reference scenario in order to avoid a pre-imposed dominance of a specific subglacial process.

However, an interesting observation for total erosion is its bimodal distribution in the accumulation zone, which is mainly driven by quarrying distribution. Finally, our small-scale results agree with glacier-scale mean erosion predictions (Figure 8), where the total eroded volume in the accumulation zone is smaller than in the ablation zone for low-precipitation scenarios, but greater in high-precipitation conditions with increased abrasion rates (Figure 9).

We explored also the cell-by-cell variability in predicted subglacial erosion for the three end-member climatic scenarios, in terms of magnitude and hypsometric distribution (Figure 10). Our results first show that erosion rates, overall of low magnitude (<1 mm/year), occur preferentially at low elevations (around 1000 m) for the low-precipitation scenario (0.25 m/year, ELA at ~1750 m). For the reference scenario (1 m/year, ELA at 1985 m), erosion rates are higher (peak above ~1.5 mm/year) and a bimodal pattern emerges, with subglacial erosion located at low elevations (around 1000 m) but also

**FIGURE 9** Subglacial abrasion (a–c), quarrying (d–f) and total erosion (g–i) rates, for three end-member climatic scenarios (left: 0.25 m/year, centre: 1 m/year and right: 4 m/year, respectively). Output rates smaller than 10^{-5} mm/year are considered as no-erosion data and are not shown. Blue and red histograms are referring to accumulation and ablation zones, respectively. Mean (black dashed line) values for the entire main glacier system (combined accumulation + ablation zones) are also displayed. The frequency was calculated based on the number of 100 × 100 m model cells within each bin of erosion rate values [Color figure can be viewed at wileyonlinelibrary.com]
close to the ELA (around 2000 m). This trend is visible in the wettest scenario (4 m/year, ELA at ~2100 m), with subglacial erosion rates up to 5 mm/year and three observed modes with an original high-elevation one (around 3000 m) in addition to the previously observed peaks at low elevations and at the ELA.

These results reinforce the cell-based erosion trends shown previously (Figures 9g–i), with an evolution towards bimodal/multimodal patterns for erosion, especially in the accumulation zone, for increased precipitation scenarios. Besides, in the wettest/warmest scenario, a wide range of subglacial erosion rates arises, with the highest rates at the steep high-elevation tributaries.

To further investigate the potential controls of ice dynamics on erosion variability, we explore the possible links with basal sliding. Abrasion is closely linked to sliding, as defined by the input power-law described in Equation 7. Likewise, the model input law for quarrying results in highly non-linear responses with basal sliding due to the important contributions of effective pressure [Equation 8], leading to different erosion patterns in the ablation and accumulation zones (Figure 11). For the accumulation zone, the relationship between quarrying and basal sliding is complex, and a triangular-like pattern is observed for all climatic scenarios (Figure 11a), although ranges increase with higher precipitation and the maximum quarrying is set by basal sliding magnitudes [Equation 8]. However, because models for sliding and quarrying both depend on effective pressure, peaks in quarrying occur around half-way on the sliding range for all three end-member scenarios (Figure 11a). In the ablation zone a clearly different trend appears, with higher dispersion and two opposite clusters for the relationship between basal sliding and quarrying (Figure 11b), especially for the highest-precipitation scenario.

Finally, we extracted from the two end-member and reference climatic scenarios spatial patterns for both abrasion and quarrying (Figure 12). A normalization was performed to provide better relative spatial comparisons. For abrasion, it is noteworthy that the lowest-precipitation scenario (Figure 12a) predicts highly distributed abrasion with several patches of high rates in the ablation zone, while overall lower rates are observed in the accumulation zone. With increasing precipitation (Figures 12b and c) it appears that small-scale abrasion patches are less frequent, while larger ones in both the accumulation and ablation zones dominate the abrasion pattern. For quarrying, this evolution with increasing precipitation is even more evident. For the lowest precipitation scenario (Figure 12d), maximum quarrying is focused in a highly clustered pattern along the main trunk valley, mostly in the ablation zone. With increasing precipitation
maximum quarrying migrates from the main valley towards the accumulation zone (in smaller lateral valleys) and the number of clusters is significantly reduced in the ablation zone.

4 | DISCUSSION

4.1 | Model simplifications and limitations

For our SA and ST simulations, we used a multi-tributary synthetic landscape to assess the impact of varying climate conditions on ice dynamics, simulating a first-order alpine context in terms of relief and hypsometric distribution (Figure 3a) with a large elevation range (0–4500 m) and typical alpine relief (topographic slopes almost null at valley bottoms and up to 40° in high-elevation areas). However, we did not investigate different topographic configurations (e.g. low-relief plateau; Egholm et al., 2017; Kessler et al., 2006) or temporally evolving landscapes (e.g. Liebl et al., 2021; Pedersen & Egholm, 2013; Sternai et al., 2013; Tomkin & Braun, 2002), which is outside the main scope of our study. Other model simplifications have been adopted for the present study, mainly for computational efficiency and in order to isolate the glacial response to different climate scenarios, which are discussed thereafter. These include modelling setup, with a single glacial-advance event simulated and no evolving landscape under repetitive glacial periods (Pedersen & Egholm, 2013), but also climatic simplifications with spatially uniform precipitation over the model extent, constant melting degree-day factor and annual temperature amplitude in the PPD model, or neglect of seasonal meltwater fluctuations (e.g. Ugelvig et al., 2018). As solar radiation, supraglacial debris is also not incorporated in our mass-balance computations, which can have an impact on glacier ablation, as observed by Nakawo and Young (1981) and Rowan et al. (2021). However, these effects are not entirely neglected in our approach, but are partially incorporated within the input value for melting degree factor (mPDD, Table 1), although we do not consider spatial variations in this parameter to account for the above-mentioned processes in a real landscape. Moreover, we considered snow avalanching over a fixed threshold (\(\text{avaslope} \geq 28°\)) to simulate how snow redistribution over steep glacier/bedrock surfaces influences surface mass balance (Scherler & Egholm, 2020), but we did not evaluate more complex snow redistribution patterns linked to other thresholds or to orography (Kessler et al., 2006; MacGregor et al., 2009).

For predicting subglacial erosion patterns, we did not consider fluvial/hillslope processes during our simulations. Subglacial erosion is treated as a mixture of two distinct processes, abrasion and quarrying (e.g. Anderson et al., 2006; Egholm et al., 2009; Hallet, 1979; MacGregor et al., 2000) which, despite their close relationship, follow separate laws in our experiments [Equations 7 and 8] and behave differently under varying climate conditions (Figure 7). Particularly, abrasion under real glaciers is likely limited by the availability of abrasive tools (e.g. Herman et al., 2021) in the form of debris provided by quarrying or hillslope processes, but this limitation was not considered by our experiments. We also reinforce that the predicted erosion rates and spatial trends should be considered as means of...
forcing multi-tributary synthetic landscape (as proposed in Bernard et al., 2006), which differs from model observations for ice-sheet sensitivity to climate forcing (e.g. Six & Vincent, 2014; Zemp et al., 2010) and 0.67/C19/.

For our SA and ST simulations, we found similar trade-offs (Figure 2c) between temperature and precipitation to maintain glacier ELA (SA) or ice extent (ST). Moreover, we observed a high ice-model sensitivity for changes in precipitation/temperature, especially towards dry/cold scenarios (<1 m/year; Figure 2d) for which important changes in glacier ELA or extent occurred. This sensitivity for very low precipitation/temperature settings is typical of alpine glacier sensitivity to climate forcing (e.g. Six & Vincent, 2014; Zemp et al., 2006), which differs from model observations for ice-sheet behaviour such as in the Antarctic Peninsula (Davies et al., 2014).

In SA simulations, lower precipitation rates or low temperatures resulted in a significant impact on glacier geometry, triggering considerable reductions in both glacier length and thickness. Our multi-tributary synthetic landscape (as proposed in Bernard et al., 2020) emphasized ice-flux increase for higher precipitations with tributary joining (Figure 5a), resulting in large glacier geometry differences (both length and thickness) between scenarios with similar ELAs (Figures 3a and b). In ST simulations, a similar effect for glacier thickness is observed, although of lower magnitude (Figure 3c, associated with a decrease in ELA as shown in Figure 2b).

The importance of ice flux (velocity and thickness) is, therefore, paramount to predict ice geometry when compared to the mass balance alone. Indeed, ELA information is not sufficient to quantitatively constrain paleo-glacier length or thickness, as illustrated from our simulations with same-ELA scenarios (SA; Figure 3b). This result confirms the general difficulty of reconstructing paleo-glacier using ELA shifts from paleo-temperature or paleo-precipitation proxies (e.g. Joerin et al., 2008; Ohmura, 2001). In contrast, our results also showed that same ice-extent simulations (ST) also resulted in a large variability in inferred ELAs (with a range of around 350 m; Figures 2b–d), although these are associated with much smaller changes in glacier thickness (Figure 3c).

Another interesting outcome from our ST simulation is the glacier shape in cross-sections and the output differences between ablation and accumulation zones (Figure 4). In the accumulation zone, different climate configurations provide most of the ice-thickness variations in the central part of the trunk glacial valley (Figures 3c–d), while our results show limited ice-thickness changes on the lateral margin of the glacier. In contrast, in the ablation zone, significant ice-thickness changes can be observed on the valley sides (Figure 4a), theoretically suggesting that ablation zones would be more suitable target areas than accumulation zones to reconstruct paleo-glacier geometry, thus inferring paleo-climate conditions (e.g. Provin et al., 2019; Seguinot et al., 2018).

Our ST simulations also allowed us to estimate the variability in accumulation area ratio (AAR; Meier & Post, 1962) for different climatic settings. AARs can vary in the literature for alpine glaciers, generally reported between 0.64 (Kern & László, 2010) and 0.67 (Gross et al., 1976; Pellitero et al., 2015). Except under very low-precipitation conditions (i.e. driest/coolest scenarios), output AAR estimates are lower for our simulations, between 0.52 and 0.60, and depend strongly on the input precipitation/temperature. Consequently, our simulations provide larger than literature-reported ablation zones, which can be attributed to different factors including the steepness of our synthetic landscape, the steady-state nature of our simulations (compared to empirical AAR measurements from the literature on transient landscape), combined with potentially overestimated ice fluxes from high-precipitation input scenarios.

4.2 | Response of glacier geometry to climatic forcing

For our SA and ST simulations, we found similar trade-offs (Figure 2c) between temperature and precipitation to maintain glacier ELA (SA) or ice extent (ST). Moreover, we observed a high ice-model sensitivity for changes in precipitation/temperature, especially towards dry/cold scenarios (<1 m/year; Figure 2d) for which important changes in glacier ELA or extent occurred. This sensitivity for very low precipitation/temperature settings is typical of alpine glacier sensitivity to climate forcing (e.g. Six & Vincent, 2014; Zemp et al., 2006), which differs from model observations for ice-sheet behaviour such as in the Antarctic Peninsula (Davies et al., 2014).

In SA simulations, lower precipitation rates or low temperatures resulted in a significant impact on glacier geometry, triggering considerable reductions in both glacier length and thickness. Our multi-tributary synthetic landscape (as proposed in Bernard et al., 2020) emphasized ice-flux increase for higher precipitations with tributary joining (Figure 5a), resulting in large glacier geometry differences (both length and thickness) between scenarios with similar ELAs (Figures 3a and b). In ST simulations, a similar effect for glacier thickness is observed, although of lower magnitude (Figure 3c, associated with a decrease in ELA as shown in Figure 2b).

The importance of ice flux (velocity and thickness) is, therefore, paramount to predict ice geometry when compared to the mass balance alone. Indeed, ELA information is not sufficient to quantitatively constrain paleo-glacier length or thickness, as illustrated from our simulations with same-ELA scenarios (SA; Figure 3b). This result confirms the general difficulty of reconstructing paleo-glacier using ELA shifts from paleo-temperature or paleo-precipitation proxies (e.g. Joerin et al., 2008; Ohmura, 2001). In contrast, our results also showed that same ice-extent simulations (ST) also resulted in a large variability in inferred ELAs (with a range of around 350 m; Figures 2b–d), although these are associated with much smaller changes in glacier thickness (Figure 3c).

From a geomorphological point of view, the different trade-offs between temperature and precipitation present a diversity of possible climate scenarios that can provide similar ELAs or ice-extent configurations (but not both simultaneously). This suggests that inferring past ELA estimates or climatic conditions from paleo-glacial reconstructions should be taken cautiously, as already indicated by several case studies (Becker et al., 2016; Blard et al., 2007; Mey et al., 2020; Provin et al., 2019). This problem can be better constrained when adding glacier thickness information to paleo-glacial reconstructions such as lateral moraine elevation estimates (e.g. Provin et al., 2019; Reixach et al., 2021). However, our ST simulations point to relatively minor changes in glacier thickness for very contrasted climatic and ELA scenarios (Figure 3c), which may limit such an approach.

Another interesting outcome from our ST simulation is the glacier shape in cross-sections and the output differences between ablation and accumulation zones (Figure 4). In the accumulation zone, different climate configurations provide most of the ice-thickness variations in the central part of the trunk glacial valley (Figures 3c–d), while our results show limited ice-thickness changes on the lateral margin of the glacier. In contrast, in the ablation zone, significant ice-thickness changes can be observed on the valley sides (Figure 4a), theoretically suggesting that ablation zones would be more suitable target areas than accumulation zones to reconstruct paleo-glacier geometry, thus inferring paleo-climate conditions (e.g. Provin et al., 2019; Seguinot et al., 2018).

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4.3 | Subglacial erosion under varying climatic scenarios

Our simulation results provide quantitative evaluation for the role of climate forcing on glacial erosion, as generally reported in the literature (e.g. Anderson et al., 2006; Cook et al., 2020; Egholm et al., 2012; Herman et al., 2011, 2015; Koppes et al., 2015). We showed the influence of precipitation/temperature (via ice flux) on mean abrasion and quarrying (Egholm et al., 2012; Figure 7), with a non-linear relationship between climate forcing and mean quarrying (Bernard et al., 2020; Ugelvig et al., 2018). However, we further expanded these investigations by differentiating accumulation and ablation zones, which have often been overlooked when considering the glacier system as a whole. While mean subglacial erosion rates are generally higher in the ablation zone for our model outcomes (Figure 7; Herman et al., 2011; MacGregor et al., 2009), our simulations reveal different relationships with climatic input for these two glacier zones. In addition, when we considered the sum of subglacial erosion (i.e. the total eroded volume), our results highlight that the accumulation zone dominates the total eroded volume for high-precipitation scenarios (>2 m/year; Figure 8a), in accordance with most studies (Cook et al., 2020). This is reinforced by the different relationships between ice and erosion volumes between the ablation (apparent exponential trend) and accumulation (no apparent trend) zones (Figure 8b). This output observation might result from the combined effects of sliding and effective pressure. With increased precipitation, ice-bed contact would be reduced in the ablation zone due to the higher water flux, which—combined with lower effective pressure—drives a slower erosion in the ablation compared to the accumulation zone. Although these mechanisms and feedbacks between ice flux, effective pressure and sliding into erosion volume deserve further investigation beyond the scope of our study, we think this observation could be important for provenance studies (e.g. Bernard et al., 2020; Enkelmann & Ehlers, 2015; Godon et al., 2013; Herman et al., 2015; Stock et al., 2006). In addition, our results show that the eroded volume by glaciers can be indicative of (paleo-)climatic conditions such as precipitation setting, since this has also been investigated for other topographic/climatic factors (Koppes et al., 2015). The dependence of subglacial erosion on precipitation is clear when subglacial erosion rates are analysed separately between ablation/accumulation zones as frequency distributions (Figure 9): a shift from low to high rates of abrasion and quarrying, mostly in the accumulation zone, can be observed for increased precipitation/temperature (which can also be noticed in Figure 12). This can be attributed to increased sliding in the steep tributaries belonging to the accumulation zone, as well as increased cavitation in the ablation which is not observed in the accumulation zone (Figures S7 and S8). In other words, the extent of cavitation (SLf parameter in the ISOSIA model; Ugelvig et al., 2016, 2018) remains very low in the accumulation zone (almost full contact), while it increases in the ablation zone and consequently decreases sliding, relative to the accumulation zone, which still shows high effective pressure (e.g. Iverson, 2012). Ice dynamics between the two zones is, thus, strongly controlled by subglacial hydrology computed using a cavity-based law in our experiments, which in turn affects sliding and depends on effective pressure (thus, ice-bed contact fraction) and shear stress. While small cavities can transport water in the accumulation zone (where ice is thicker and less water is available), channels and more widespread cavitation dominate the ablation zone and the latter accelerates basal sliding. The combined effects of subglacial hydrology, terrain steepness and climate inputs (i.e. precipitation/melting in our model setup) provide notable differences in ice thickness and velocities between the two zones and, consequently, subglacial erosion distributions (Figure 9).

We further investigated the differences in ice dynamics between the ablation and accumulation sectors, by exploring the apparent output relationships between quarrying and sliding (Figure 11). Both abrasion and quarrying are sensitive to sliding and hydrology. Ice-bed contact fraction and basal shear stress play a main controlling role in abrasion, through sliding (Equation 2). While abrasion vs. sliding is linked in the described power-law (Equation 7), quarrying also depends on the effective pressure (therefore, water pressure) and bed slopes (Equation 8), which provides interesting non-linear behaviour with increased precipitation (Figure 11). While abrasion peaks with maximum sliding for both accumulation and ablation zones, we reported two different relationships that appear for quarrying vs. sliding (see Figure 11). Such behaviours may be linked to effective pressure and the likelihood of cavitation rising as ice flux increases (e.g. Iverson, 2012), disrupting the straight link between quarrying and sliding for high-precipitation scenarios. This is mostly observed in the ablation zone (Figure 11b), where channels form in response to abundant water and melting, decoupling the ice from the bed (high SLf) and allowing a decrease in quarrying even for fast sliding conditions.

In our simulations, we have illustrated, from a specific model setup over a synthetic landscape, a strong climatic control on subglacial erosion through varying precipitation and temperature, affecting both subglacial hydrology and glacier ice flux. Other topographic, climatic and geodynamic controls can influence the ice flux and subglacial erosion. Lai and Anders (2020, 2021) have shown that the glacier basal thermal regime is a key control on subglacial erosion, either from geothermal heat flux (Lai & Anders, 2020) or from increased precipitation in high-elevation areas that can result in a shift in basal thermal regimes from cold to warm-based, impacting erosion magnitudes and patterns (Lai & Anders, 2021). Although we considered only warm-based glaciers in our ISOSIA simulations, the modelled outcomes agree with previously reported observations and interpretations from the literature, with subglacial erosion increasing for high-precipitation scenarios and peaking around the ELA (Figure 10). Similar results were observed in SA simulations (Figure S4). We did not further investigate the potential influence of basal thermal regime on subglacial erosion in our simulations. However, due to the steepness of our synthetic landscape, with only around 18% of modelled ice above 2500 m for the reference scenario (where a cold-based glacier can occur during full alpine glacial conditions; e.g. Cohen et al., 2017), warm-based ice controlled most of the ice dynamics in our numerical simulations.

4.4 | Spatial patterns of subglacial erosion

From our numerical investigations, we noticed distributed patches of subglacial erosion in all simulations, notably for low-precipitation/temperature scenarios. These erosion patches occurred mainly at tributary confluences and in the main trunk valley centre,
preferentially in the ablation zone (Figures 10a–g). We attributed such patterns in the ablation zone to high effective pressure, low cavitation (intensifying erosion where ice and bed are in direct contact) and increased sliding. For increased precipitation/temperature and higher ice-flux conditions, these patches became increasingly connected, especially for abrasion rates; such large-scale behaviour observed for natural glacial systems was suggested as a potential process in overdeepenings evolution (Magrani et al., 2020; Patton et al., 2016). While subglacial erosion patches become connected with increased precipitation/temperature in the ablation zone, we also observed that both quarrying and abrasion increased in the accumulation zone (Figures 12c–f) within all high-elevation tributaries.

Although changes in ice geometry are considerable for same-ELA (SA) simulations, the spatial patterns of subglacial erosion are similar to same ice-extent (ST) simulations (Figures S6 and 12, respectively) with varying input precipitation/temperature conditions. Only the magnitudes of subglacial erosion changed significantly due to ice-flux differences between SA and ST simulations: output maximum erosion rates range between 0.5 and 4.6 mm/year (SA) and between 0.7 and 3.8 mm/year (ST) for end-member precipitation scenarios, respectively.

For all our climatic scenarios, subglacial erosion peaks below the ELA in the ablation zone (Figures 10–12), which highlights the role of subglacial hydrology in erosion dynamics (e.g. Beaud et al., 2016; Herman et al., 2011) compared to maximized glacier sliding and erosion near the ELA (MacGregor et al., 2000). It is also worth noting for paleo-climate reconstructions that estimated ELAs using the AAR method (instead of direct output from the mass-balance model) would potentially provide differences, altering the observed erosion patterns. Therefore, a direct relationship between estimated (paleo-)ELAs and subglacial erosion patterns is debatable, as also pointed out by Lai and Anders (2021), relying much more on ice flux and climate inputs. Interestingly, for increased precipitation/temperature and ice-flux conditions, we observed a rising multimodal (bi- to trimodal) elevation distribution, with subglacial erosion peaks located below, around and above the ELA (Figure 10). This is partially controlled by our input topography, influencing erosion (linked to sliding, basal shear stress and effective pressure) as analysed by Pedersen et al. (2014). The observation of high-elevation subglacial erosion is in agreement with recent literature (Lai & Anders, 2021), highlighting that larger ice flux leads to enhanced erosion in the accumulation area where the ice is tightly coupled to bedrock in the absence of widespread cavitation. Such erosion dynamics would have strong implications over repetitive Quaternary glaciations, since subglacial erosion at high elevations would lead to topographic changes (Sternai et al., 2013) with subsequent feedbacks on ice dynamics by reducing the glaciarized area at the accumulation zone and thus limiting erosion (e.g. Egholm et al., 2012; Pedersen & Egholm, 2013), seen as a self-defeating mechanism (Kaplan et al., 2009; MacGregor et al., 2000).

5 | CONCLUSIONS

Our numerical simulations using the iSOSIA ice-flow model with a multi-tributary synthetic landscape provided a quantitative assessment of glacier geometry and subglacial ice dynamics changes under varying climate conditions (both temperature and precipitation), with the overall aim to improve our understanding of modern to paleo-glacial systems in alpine settings.

Following our model assumptions and setup, simulation outcomes point to an important role exerted by the subglacial hydrology, together with ice flux, as controlling subglacial erosion patterns. While abrasion, linked to sliding, increases with precipitation/temperature, quarrying depends non-linearly on sliding and this dependence weakens with increased ice flux. We also illustrated, with our simulated outcomes, that mean rates are not fully representative of the overall distribution of subglacial erosion and may hinder large variability between the accumulation/ablation zones as well as at local scale (i.e. model cell in our approach). Therefore, differences in ice flux, geometry and subglacial erosion between the ablation and accumulation zones can be significant and should not be considered together as a whole. Such observation is particularly relevant for paleo-climatic studies, since the glacier ablation zone appears more suitable as a target for paleo-glacier geometry reconstructions than the accumulation zone. Our simulation results also suggest that paleo-ELA information might not be sufficient to quantitatively constrain paleo-glacier geometries, therefore empirically scaling subglacial eroded volume to ice volume alone should be taken cautiously.

Moreover, the different subglacial processes in action (i.e. cavities vs. channels) and the effect of effective pressure have an important role in erosion rates and patterns between the accumulation and ablation zones. Such behaviours might suggest that cold and dry conditions would result in more irregular and rugged glacial landscapes, while smoother landscapes may evolve under increased ice-flux conditions (i.e. wet and warm). The evolution of spatial patterns for subglacial erosion, from localized patches to larger areas, and preferentially in the ablation zone are important to consider for glacial valley evolution and possible development of glacial overdeepenings, as highlighted in previous studies focusing on natural glacial systems.

Our numerical experiments also highlighted the importance of the calibration model parameters, since modelled erosion rates and patterns rely highly on the input laws and model setup. As a consequence, some of the observed outcomes may be overly constrained or driven by our model simplifications; implying that field and laboratory measurements for constraining modelled parameters are required and would provide extremely valuable constraints for upcoming models.

Future studies, using more complex or realistic landscapes and coupling subglacial processes with hillslope/fluvial erosion, can significantly improve our modelled outcomes, when also considering subglacial sediment transport/deposition, coupling abrasion and quarrying via debris production, or by incorporating different thermal regimes over several glacial cycles. Finally, our model results have pointed out that understanding subglacial erosion magnitudes and patterns could provide a new proxy for paleo-climatic inferences from (paleo-)glacier systems.

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CONFLICT OF INTEREST
The authors declare the absence of any conflict of interest.

AUTHOR CONTRIBUTIONS
Fabio Magrani and Pierre Valla: Conceptualization, methodology, investigation and writing. David Egholm: Software (model), supervision and writing.

DATA AVAILABILITY STATEMENT
All data are available on request.

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