Timing and Climatic-driven Mechanisms of Glacier Advances in Bhutanese Himalaya during the Little Ice Age

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Abstract. Mountain glaciers provide us a window into past climate changes and landscape evolutions, but the pattern of glacier evolution at centennial or suborbital timescale remains elusive, especially in monsoonal Himalayas. We simulated the glacier evolution in Bhutanese Himalaya, a typical monsoon influenced region, during the Little Ice Age (LIA) using the Open Global Glacier Model (OGGM) driven by six paleo-climate datasets and their average. Compared with geomorphologically-mapped glacial landforms, the model can well capture the patterns of glacier length change, but underestimates its amplitude. Simulation results revealed four glacial substages (1270s, 1470s, 1710s, and 1850s) during LIA in the study area. Statistically, a positive correlation between the number of glacial substages and glacier slope was found, indicating the occurrence of glacial substages might be a result from heterogeneous responses of glaciers to climate change. Summer temperature dominates the regional glacier evolution during the LIA.

1 Introduction

Mountain glaciers over high Himalayas provide us a critical window to explore the linkage between climatic, tectonic, and glacial systems (Oerlemans et al., 1998; Owen et al., 2009; Dortch et al., 2013; Owen & Dortch, 2014; Saha et al., 2018). Many scientists have investigated the glacial history for Himalayas at orbital-scale, indicating that a general trend of glacier advances is related to overall summer temperature, forced by orbitally-controlled insolation (Murari et al., 2014; Yan et al., 2018, 2020, 2021). However, latest observations with finer temporal resolution have revealed that the evolution of some glaciers in monsoonal Himalayas has suborbital-scale fluctuations, which has aroused more and more interest on exploring the mechanisms behind (Solomina et al., 2015; Peng et al., 2020).

The Little Ice Age (LIA; from 1300 to 1850 CE; Grove, 2013; Qureshi et al., 2021) is the latest cooling event during the Holocene, during which most mountain glaciers advanced, forming abundant well-preserved and distinctive geomorphic landforms (Murari et al., 2014; Qiao & Yi, 2017; Peng et al., 2019, 2020). Previous studies have reconstructed the timing and extent of glacier evolution during the LIA based on field investigation, geomorphological mapping, and cosmogenic nuclide dating (Owen & Dortch, 2014 and references therein; Zhang et al., 2018a, 2018b; Carrivick et al., 2019; Qureshi et al., 2021). However, it is still unclear how many substages (glacial advances) exist during the LIA (Yi et al., 2008; Murari et al., 2014;

Xu & Yi, 2014), due to the post glacial degradation and the large uncertainties in the dating methods (Heyman et al., 2011; Fu et al., 2013). In addition, Carrivick et al. (2019) indicated that the reconstructions using individual glaciers or a small number of glaciers may be not representative for the regional average.

Numerical glacial modelling is a powerful way to study glacier evolution on centennial timescale (Parkes & Goosse, 2020) and quantify the response of glaciers to climate change (Eis et al., 2019). It can largely alleviate the limitations in field-based methods and is able to capture the glacier evolutions on regional scale. Meanwhile, the model simulations can be evaluated via multiple observations to ensure the reliability. However, evaluating the simulation results is still challenging due to the scarcity of the direct observational record for glacier changes during the LIA (Goosse et al., 2018).

Based on the above issues, this study provided a possible approach on how to bring observation and simulation together, what the contribution of individual glacier to regional glacier evolution is, and how climate change drives glacier evolution (Goosse et al., 2018; Carrivick et al., 2019; Peng et al., 2019, 2020). We chose a typical monsoon-influenced area, Bhutanese Himalaya (BH) as an example, using the Open Global Glacier Model (OGGM) to improve our understandings on the pattern of LIA glacier changes (Fig. 1). The BH (27.5~28.3°N, 89.1~91.0°E) is an east-west-trending mountain range with an average elevation above 5000 m above sea level (a.s.l.), nourishing abundant high mountain glaciers (Peng et al., 2019, 2020; Fig. 1b). According to the Randolph Glacier Inventory V6.2 (RGI; RGI Consortium, 2017), there exist 803 modern glaciers in BH, covering an area of ~ 1233.685 km^2 (Fig. 1b). Fifty-seven glaciers belong to RGI13 region (Central Asia) and 746 glaciers belong to RGI15 region (South Asia East). The distribution of glacier length is shown in Fig. 1c with an average length of 1596 m (950 m for the median value) ranging from 135 m to 20011 m. The small glaciers (length shorter than 3000 m) are prevalent in BH (accounting for 88.9 %).

We systematically simulated the BH glacier changes during the LIA based on the climate data from six different general circulation models (GCMs) and their average. The simulated glacier length changes are validated by geomorphological maps and previous studies. The pattern of regional glacial evolution is compared with ^{10}Be and ^{14}C glacial chronologies across the monsoon influenced Himalayas. The dominant climatic factors of BH glacial evolution are explored through analyzing the glacier surface mass balance (SMB) changes and a series of sensitivity experiments.

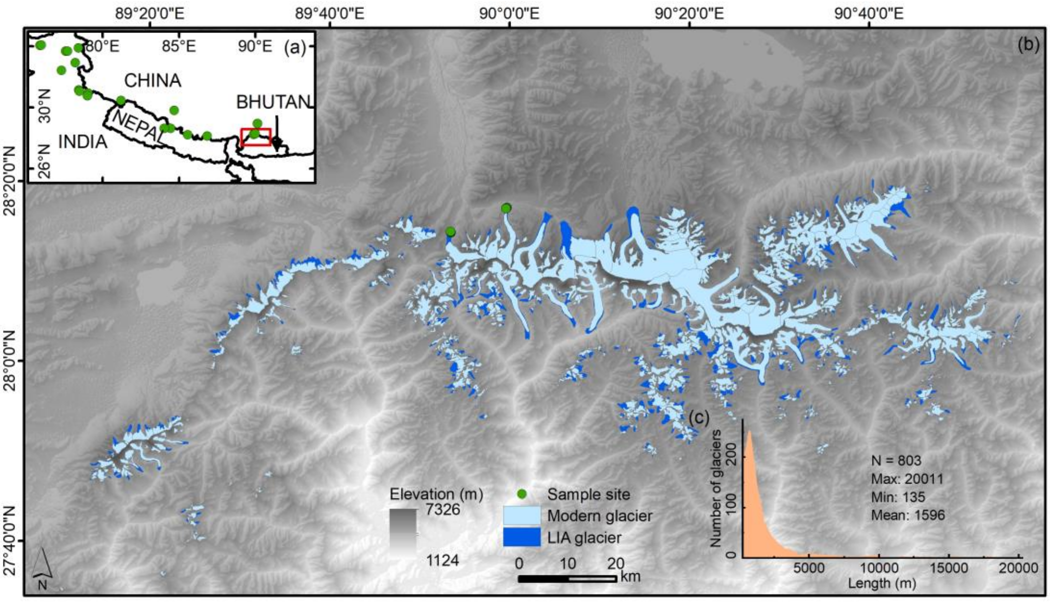

Figure 1. An overview of study area and moraine sites. The red box in (a) shows the location of the study area and the green circles in (a) displays the spatial distribution of the ^{10}Be exposure dating moraines. The basic information of these moraine sites can refer to Table S1. (b) The extent of the modern glaciers (in light blue; RGI Consortium, 2017) and LIA glacier (in navy blue). The background DEM is obtained from the Shuttle Radar Topography Mission (SRTM) 90 m Digital Elevation Model v4.1 (Jarvis et al., 2008; http://srtm.csi.cgiar.org/). (c) The length distribution of modern glaciers.

2 Methods

2.2 Model Description

The OGGM (v1.50) is a 1.5D ice-flow model, able to simulate past and future mass-balance, volume, and geometry of glaciers (Maussion et al., 2019). Previous studies have confirmed a good performance of this model in simulating alpine glaciers (Farinotti et al., 2017; Pelto et al., 2020) and reproducing the millennial trend of glacial evolution in mountainous regions (Goosse et al., 2018; Parkes & Goosse, 2020). For example, OGGM has been successfully applied to simulate High Mountain Asia glaciers, including their thickness, velocity, and future evolutions (Dixit et al., 2021; Pronk et al., 2021; Shafeeque & Luo, 2021; Furian et al., 2022; Chen et al., 2022).

The OGGM couples a surface mass balance (SMB) scheme with a dynamic core (Marzeion et al., 2012; Maussion et al., 2019). The dynamic core adopts the shallow-ice approximation (SIA), computing the depth-integrated ice flux of each cross-section along multiple connected flowlines diagnosed by a pre-process algorithm (via geometrical centerlines). Two key parameters, the creep parameter A and the sliding parameter f_s, in the dynamic core are set to their default values ($A = 2.4 \times 10^{-24}\ s^{-1}\ Pa^{-3}, f_s = 0\ s^{-1}\ Pa^{-3}$, without lateral drag). The spatial resolution (dx; m) of the target grid is scale dependent, determined by the size of the glacier ($dx = 14\sqrt{S}$, with S representing the glacier area in km^2) but truncated by minimum (10 m) and maximum (200 m) values, respectively (Maussion et al., 2019). According to the observations, the largest simulation domain is set to 160 grid points outside the modern glacier boundaries to ensure that the domain is large enough for the LIA glaciers. If a glacier advance exceeding the domain during the simulation, we will exclude this glacier in the further analysis due to its large simulation bias.

The ice accumulation is estimated by a solid precipitation scheme to separate the total precipitation into rain and snow based on monthly air temperature. In this scheme, the amount of solid precipitation is computed as a fraction of the total precipitation. Specially, precipitation is entirely solid if $T_i <= T_{Solid}$ (default setting is 0 °C), entirely liquid if $T_i >= T_{Liquid}$ (defaults to 2 °C) or divided into solid and liquid parts based on a linear relationship with those two temperature values. The ablation is estimated using a positive degree-day (PDD) scheme (Eq. 1). Melting occurs if monthly temperature ($T_i(z)$) is above T_{melt}, which is equal to -1 °C.

$$m_i(z)=p_f P_i^{solid}(z)-\mu^* \cdot max(T_i(z)+\beta-T_{melt},0) +\varepsilon, \qquad (1)$$

where $m_i(z)$ is the monthly SMB at elevation z of month i. $P_i^{solid}(z)$ is the monthly solid precipitation, and P_f is a general precipitation correction factor (default setting is 2.5). μ^* is the temperature sensitivity parameter and β is the temperature bias. A residual bias term (ε) is added as a tuning parameter to represent the collective effects of non-climate factors (Marzeion et al., 2012; Maussion et al., 2019). Different from the conventional PDD schemes embedded in other ice sheet models, such as Parallel Ice Sheet Model (Bueler & Brown, 2009; Winkelmann et al., 2011), SICOPOLIS (Greve, 1997a, 1997b) or CISM (Lipscomb et al., 2019), that assume ε and μ^* as constant values, these parameters vary with glacier in OGGM.

The monthly temperature and precipitation from six different GCMs (BCC, CCSM4, CESM, GISS, IPSL, and MPI), covering a period from 850 CE to 2000 CE, are used to drive OGGM. These data are available in the Past Model Intercomparison Project (PMIP3) and the Coupled Model Intercomparison Project (CMIP5) protocols (Schmidt et al., 2011; Tayloret al., 2012; PAGES 2k-PMIP3 group, 2015) – with details listed in Goosse et al. (2018) and Table S2. The climate data cannot be directly used in glacial model due to the large systematical bias of GCMs. A calibration algorithm is adopted by OGGM to correct the GCMs climate data by taking the anomalies between GCMs and the Climate Research Unit (CRU) TS 4.01 (Harris et al., 2020) mean climate from 1961 to 1990 (Parkes & Goosse, 2020). In addition, the mean climate (MC) from six different GCMs is also calculated and calibrated to drive OGGM (hereafter MC experiment) to further alleviate the climate bias of each GCM. Therefore, we would focus on analyzing the results from MC experiment, but also involve some discussions on the difference between MC experiment and six GCM experiments.

2.2 Identification of the Glacial Substages and Related Concepts

Similar to Goosse et al. (2018) and Parkes & Goosse (2020), we use simulated glacier length change ($\Delta L = L - L_{1950}$, where L_{1950} represents the simulated glacier length at 1950) to represent glacier evolution. In order to alleviate the influence of glacier size (length) to the mean value, we further convert ΔL into *glacier length change ratio* ($GLR = \frac{\Delta L}{L_{1950}}$). Firstly, we exclude the glaciers of which the simulated lengths equal to zero at 1950 because these glaciers have large simulation biases according to the observations (RGI). Then, decadal mean GLR is calculated for each glacier in order to remove the interannual variabilities. Next, the Gaussian Filter (with standard deviation setting to be 3) is applied to the decadal mean GLR for each glacier to extract the main oscillations. After that, we obtain the regional average GLR by averaging all glaciers' GLR (decadal averaged and Gaussian Filtered) within the domain. Finally, we try to find all peaks and their corresponding times in the regional average GLR timeseries based on the "findpeaks" function (with the minimum peak prominence is set to 0.2) embedded in Matlab Software. Each peak found is defined as a *glacial substage* during the LIA. We name the substages from new to old (LIA-1, LIA-2, LIA-3, LIA-4 and maybe more).

A concept related to GLR is *maximum peak GLR*, defined as the GLR when a glacier reaches its maximum peak during a period. Notice that *maximum peak GLR* is different from the maximum GLR. For example, in Fig. 2d, the *maximum peak GLR* occurs around 1270 CE rather than 1100 CE. Based on this concept, the simulated *second/third/fourth peak GLR* is defined as the GLR when a glacier reaches it second/third/fourth maximum peak during a period.

2.3 Spin-up, Tuning Strategy, and Experiment Design

We spin-up the model to avoid the influence of the initial condition and tuned the parameter, temperature bias (β) in Eq. 1, to obtain a better initial condition. The β will directly regulate the initial condition (i.e., the length of initial glaciers) and largely impact the *GLR* during early LIA (e.g., LIA4). Continuously forced by the climate data selected randomly from a 51-year window of 875 - 925 CE, all experiments can reach their steady states after a 5000-years spin-up. In order to archive a proper initial condition, we alter β from -1 to 1 °C with an increment of 0.1 °C during the spin-up period. After spin-up, we model the LIA glacier changes with $\beta = 0$, forced by the past climate time series from 900 to 2000 CE. In addition, we start our analysis at the year 1100 for a better display of the glacial fluctuations during the LIA (1300-1850 CE; Grove, 2013; Qureshi et al., 2021).

The tuning procedure is based on MC experiment while six GCM experiments share the same β with MC experiment during spin-up period. Our tuning strategy is threefold. First, we should ensure the regional average *GLR* is larger during LIA4 than LIA1 as in the observations because previous studies indicated that the majority of glaciers advanced to their LIA maximum extents at the early LIA rather than the late LIA (Murari et al., 2014; Xu & Yi, 2014). Second, we need to ensure the simulated *maximum peak GLR* closer to the observations. Notice that we choose to use *maximum peak GLR* because the observations derived from the geomorphological mapping methods can only obtain this variable during LIA (Section 2.4). Third, let more glaciers be available in the analysis as a smaller β will decrease the number of available glaciers (Fig. 2c).

A series of sensitivity experiments are also conducted to further validate the effect of climate changes on BH glacier advances on both seasonal and annual scales. We apply a 'constant climate scenario', using the CRU datasets as the climate forcing, and run the simulation until reaching equilibrium (here 5000 years). The window size is set to 51-year and centered on t^*, the year for which the SMB scheme best reproduces the observed SMB. We set ε to 0 in Eq.1 in order to maintain the contemporary glacier geometry under the contemporary climate condition. The control experiment is forced by the default monthly temperature and precipitation. Keeping the same precipitation, we add a temperature bias from -1 to 1 ℃ with an increment of 0.1 ℃ to the original seasonal/annual temperature to test the sensitivity of temperature on glacier evolution. The similar approach is also applied to the precipitation. Holding the temperature, we adjust the precipitation from -20 to 20 % with an increment of 2 % in the original seasonal/annual precipitation data.

2.4 Establishing Regional Chronology and Mapping LIA Glacier

The simulated timing and extent of glacial advances are validated with the ^{10}Be surface exposure ages and ^{14}C ages of the LIA moraines and the mapped LIA glaciers over BH. Seven ^{14}C ages in monsoonal Himalaya are derived from Xu & Yi (2014) and all ^{10}Be ages are recalculated using CRONUS Earth V3 online calculator with the time and nuclide-dependent scaling scheme 'LSDn' (Balco et al., 2008; Lifton, et al., 2014; http://hess.ess.washington.edu/math/). We then adopt the method advocated by Chevalier et al. (2011) and Dong et al. (2018) to exclude the potential outliers. The potential outliers are defined as the ^{10}Be ages which did not overlap within 1 σ external uncertainty with others for a moraine. After removing outliers, we use the oldest age of a moraine sample set to represent the moraine depositional age (Chevalier et al., 2011; Dong et al., 2018; Peng et al., 2020). Five ^{10}Be ages from moraine M1 of Cogarbu valley and seven ^{10}Be ages from moraine M1 of Shi Mo valley were selected to determine the regional glaciation chronology establishing in BH (Fig. 1b and Fig. S1). In addition, we also chose 126 ^{10}Be surface exposure ages and 7 ^{14}C across the monsoonal Himalayas as a supplement (Fig. 1a; Table S1; Xu & Yi, 2014).

Based on regional glacial chronology and the evidence of sediment-landform assemblages (Chandler et al., 2019), we map the outermost lateral and terminal moraines in BH to represent the maximum extent of glaciers during the LIA (the *maximum peak GLR*). These moraines are usually well-preserved with sharp crests, locating from several hundred meters to a few kilometers away from the termini of modern glaciers, and damming a lake in front of modern glaciers (Qiao & Yi, 2017; Zhang et al., 2018b; Qureshi et al., 2021). We use the world imagery ESRI (http://goto.arcgisonline.com/maps/World_Imagery) and Google Earth high-resolution imagery to delineate the LIA moraines and outlines. However, not all LIA glaciers could be identified due to the destruction of moraines. Only 408 glaciers of the 803 BH glaciers could be mapped (Fig. 1b). The length of contemporary glaciers is provided in Randolph Glacier Inventory V6 datasets (RGI; RGI Consortium, 2017), and that of the LIA glaciers is calculated in ArcGIS based on the main model flowline in OGGM.

3 Results

3.1 The Choice of Initial Condition

In order to obtain a better estimation of the initial condition, we tuned the β during the spin-up period. As shown in Fig. 2, β strongly influences the initial condition and thus the LIA simulation results, especially for the first 600 years (Fig. 2b). With a decreased β, the regional average glacier volume increases (Fig. 2a), but the number of available glaciers (i.e., glaciers which do not exceed the prescribed domain boundaries) decreases during the spin-up period (Fig. 2c). The number of available glaciers for the LIA simulation is approximately equal to that during the spin-up period, except for a reduction when β is positive (Fig. 2c). This is probably because smaller β can kick out the glaciers which would potentially suffer from large simulation bias during LIA simulation. In addition, more glaciers disappear in 1950 ($Length_{1950} = 0$; Fig. 2c) with a larger β, consistent with our knowledge.

Initial condition slightly impacts the time and number of glacial substages but largely influences the strength of glacial substages (GLR) during LIA simulation (Fig. 2d; Fig. S1). Four substages occurred at ~1270s (LIA-4), ~1470s (LIA-3), ~1710s (LIA-2) and ~1850s (LIA-1) are detected under a wide range of β (from -0.7 °C to 1.0 °C) in the MC experiment. However, the number of substages become less when β is smaller than -0.7. This is because smaller β would cause excessively large initial glaciers so that a smaller climate perturbation is not powerful enough for the glaciers to stop retreating during the early LIA period.

The GLR during the early LIA periods (LIA-4 and LIA-3) are strongly regulated by the initial condition (Fig. 2d). Smaller β will lead to a larger GLR during LIA-4 and LIA-3. According to the tuning strategies in Section 2.3, simulations with $\beta \geq -0.3$ should be excluded as larger GLR must be ensured during LIA-4 than LIA-1. The Root Mean Squared Error (RMSE) of *maximum peak GLR* between the simulation and observation is smallest when $\beta = -0.4$ (RMSE = 133.1 %), though a decreasing trend is found when $\beta \leq -0.8$ (Fig. 3). However, the number of available glaciers when $\beta \leq -0.8$ is less than that when $\beta = -0.4$. Therefore, we finally choose the simulation results with $\beta = -0.4$ based on the tuning strategies.

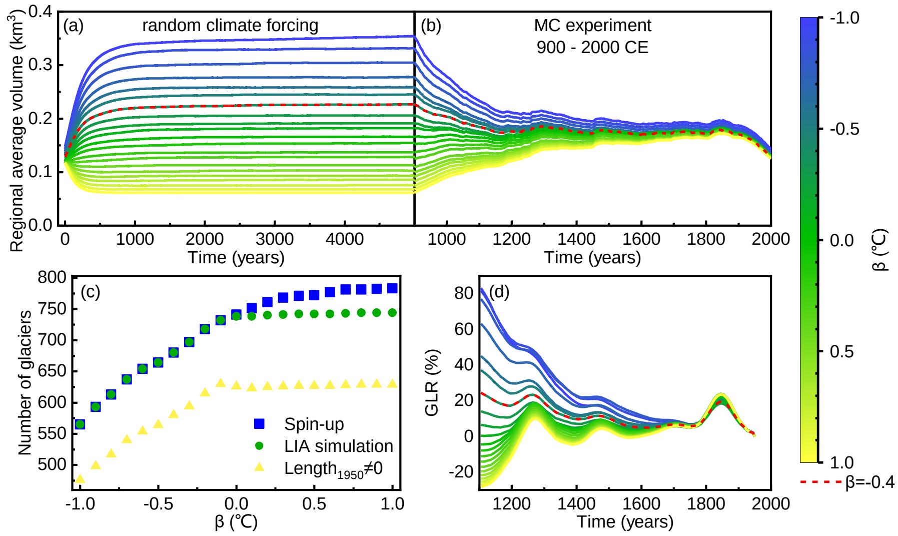

Figure 2. (a) The regional average glacier volume during the 5000-year spin-up with various β. (b) The simulated regional average glacier volume from 900 to 2000 CE with different initial condition. (c) The number of available glaciers with various β. (d) The simulated regional average *GLR* from 1100 to 1950 CE.

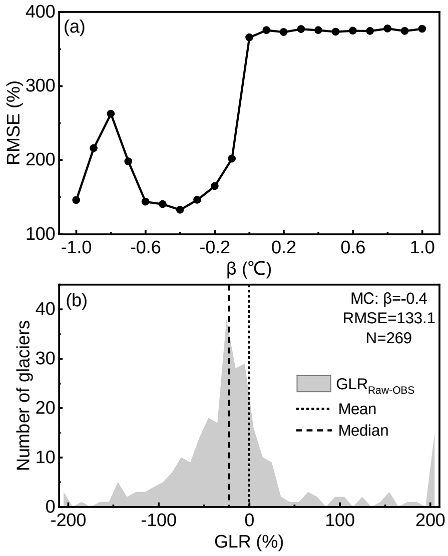

Figure 3. (a) The RMSE of *maximum peak GLR* between the raw simulation results and mapped LIA glaciers for the MC experiment with various *β*. (b) The simulation bias distribution of *maximum peak GLR* for the MC experiment with $\beta = -0.4$.

3.2 The pattern of glacier changes during the LIA

We would focus on the pattern of glacier changes during the LIA in MC experiment, but six GCM simulations are also shown in Fig. 4a for comparison. The simulation results in most experiments indicate four LIA glacial substages in BH, except for the CESM experiment losing LIA-3 substage. The timings of the four LIA glacial substages are 1270s (LIA-4), 1470s (LIA-3), 1710s (LIA-2), and 1850s (LIA-1) in MC experiment. These times vary slightly among the six GCM experiments, around 1230s - 1320s, 1470s - 1520s, 1620s - 1730s, 1800s - 1850s, respectively.

The most extensive glaciers occurred during LIA-4 in MC and six GCM experiments because our tuning strategy is to ensure larger the regional average *GLR* at the early LIA. The *second peak GLR* occurred during LIA-1 in MC experiment, similar to the results in the CCSM4, GISS, and MPI experiments but different from the results in BCC (LIA-2), CESM (LIA-2), and IPSL (LIA-3) experiments. The *third* and *fourth peak GLR* occurred during LIA-3 and LIA-2 respectively in MC experiment, also consistent with the simulations forced by CCSM4, GISS and MPI climate datasets.

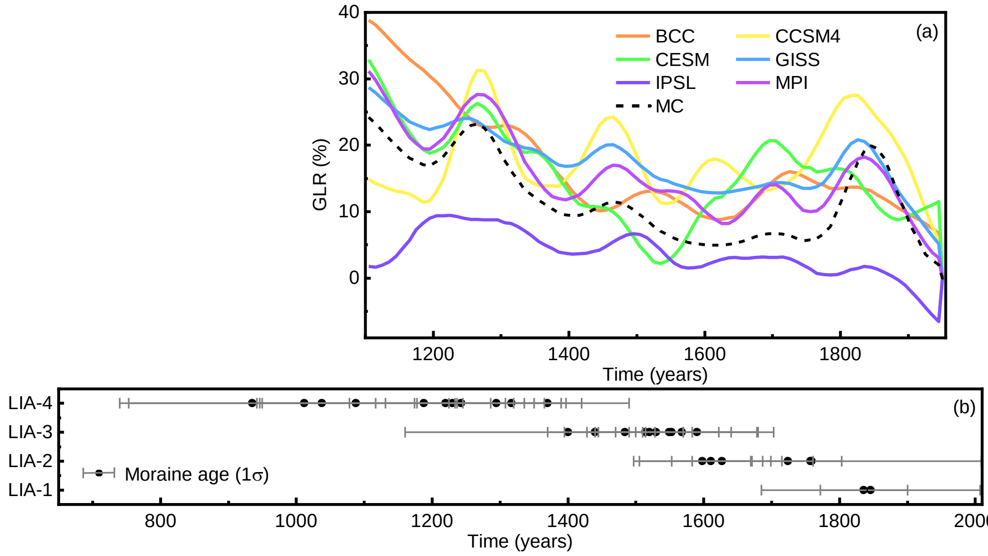

Figure 4. (a) Time series of regional average *GLR* from 1100 to 1950 CE. (b) The observational timing when glaciers in the monsoonal Himalaya reached their *maximum peak GLR*. We grouped the moraine ages based on their temporal distances to each glacial substage simulated in MC experiment. The detailed information of the moraine ages measured by ^{10}Be and ^{14}C can be found in Table S1 and Xu & Yi (2014), respectively.

4 Discussions

4.1 Comparison between Simulations and Observations

We validated the simulation results using the moraine ages and mapped LIA glaciers (Section 2.4). The simulated regional average *maximum peak GLR* (59.9%; Fig. 3b) in MC experiment agrees well with that of mapped glaciers (60.6 %). Similarly, the simulation results in BCC (57.8 %), GISS (58.9 %) and MPI (71.0 %) experiments are also consistent with observations. Observations from adjacent regions also support the simulation results (Qiao % Yi, 2017; Zhang et al., 2018b). For example, Qiao & Yi (2017) found that the *maximum peak GLR* increased about 53.8 % during LIA in the central and western Himalayas relative to 2015. Zhang et al. (2018b) reported a 71.5 % increasement of *maximum peak GLR* during LIA in the Gangdise Mountains relative to 2010, based on the glacial geomorphological maps. However, the CCSM4 (99.0 %) and CESM (80.8 %) experiments overestimated the *maximum peak GLR* while the IPSL (32.1 %) experiment underestimated it (Fig. S1). In addition, the negative bias for the median value in the simulations compared with observations is identified in the MC and six GCM experiments (Fig. 3b and Fig. S1). The difference between the mean value and median value indicates some extremums might impact the average.

The occurrence time of the *maximum peak GLR* is during LIA-4 in MC experiment, consistent with the moraine ages that the majority of glaciers advanced to their LIA maximum extents at the early LIA rather than the late LIA (Fig. 4b; Murari et al., 2014; Xu & Yi et al., 2014). Specifically, about 12 of the 30 moraine ages shows that the related glaciers reached their *maximum peak GLR* during LIA-4 compared with only 2 of them during LIA-1. However, there are still a large number of glaciers reaching their *maximum peak GLR* during LIA-3 (about 10 glaciers) and LIA-2 (about 6 glaciers). Ignoring the large uncertainties in the dating methods, the collective and individual differences in glacier changes are worth exploring. We will further discuss this issue in Section 4.2.

The simulated number of LIA substages is also comparable with observations, including some moraine dating results and climatic proxy records. For example, Murari et al. (2014) and Zhang et al. (2018a) have identified four LIA moraines in Bhillangana and Dudhganga valleys, Garwal Himalaya, and Lopu Kangri Area, central Gangdise Mountains, respectively. Liu et al. (2017) have found at least three LIA moraines in Lhagoi Kangri Range, Karola Pass. Yang et al. (2003) found four cold phases during AD 1100-1150, 1500-1550, 1650-1700, and 1800-1850 over TP and eastern China according to the proxy data of paleoclimate. A regional moraine chronologies framework composed of ^{14}C, lichenometry, and cosmogenic radionuclide ages found three substages during late-14th, 16th to early-18th, and late-18th to early-19th, corresponding to LIA-3, LIA-2, and LIA-1, respectively (Xu & Yi, 2014). However, the divergent number of LIA substages were also confirmed by some dating results and records. For example, only one moraine was dated in Cogarbu valley (1484 ± 44 CE; Table S1; Peng et al., 2019) and Shi Mo valley (1514 ± 69 CE; Table S1; Peng et al., 2020), but two substages were constrained in Lato valley, Lahul Himalaya (Saha et al., 2018), Langtang Khola valley, Nepal Himalaya (Barnard et al., 2006), and Gongotri Ganga valley, Garhwal Himalaya (Barnard et al., 2004). By applying dendroglaciology approach, Hochreuther et al. (2015) and Bräuning (2006) only detected one LIA substage in Gongpu glacier, Zepu glacier, Baitong glacier and Gyalaperi glacier, while more substages were found in Lhamcoka glacier (Bräuning, 2006), Xinpu glacier (Hochreuther et al., 2015), Gangapurna glacier, and Annapurna III glacier (Sigdel et al., 2020). Yi et al. (2008) identified three substages during AD 950-1820 based on 53 ^{14}C dating ages.

4.2 Why do four LIA substages exist in BH?

Clearly, MC experiment and GCM experiments (excluding CESM experiment) indicate four glacial substages over BH during LIA. However, due to the glacier individualities (different slopes and lengths), this does not mean each glacier in our study area exists four LIA substages (Fig. 5a), consistent with the moraine dating results. Instead, it just reflects that the majority of glaciers in BH have four glacial substages. For example, in MC experiment, only about 33.8 % glaciers have four substages during the LIA, while the rest glaciers are with zero (4.0 %), one (15.5 %), two (17.9 %), three (26.6 %) and five (2.2 %) substages. We argue that the difference in LIA substages is caused by the sensitivity of different glaciers despite many studies ascribed it to the different climate conditions (Owen & Dortch, 2014; Murari et al., 2014; Saha et al., 2019). Analysis found the number of glacial substages are significantly correlated to the glacier properties (glacier length and slope). The glacial substage numbers have significantly positive correlation with the glacier slopes while obviously negative correlation with the glacier length (Fig. 5b, c). The correlation coefficient (CC) between the number of glacial substages and glacier length at 1950 is -0.31 and the CC between the number of glacial substages and glacier slope at 1950 is 0.41. Both of the CCs can pass 95% significant test. However, when zooming into the main glacial substages numbers (2, 3, 4), the relationship between the number of glacial substages and glacier length does not become that clear (Fig. 5b). Therefore, we argue that glacial slope may dominate the glacial substage numbers during LIA (Lüthi, 2009; Zekollari and Huybrechts, 2015; Bach et al., 2018; Eis et al., 2019). The negative correlation between the glacier length and glacial substage numbers might be a result of the fact that the longer (larger) glacier has a smaller slope (CC = -0.50). Besides, analysis also suggests weak relationship between glacial substage numbers and glacial ELA (Fig. 5d).

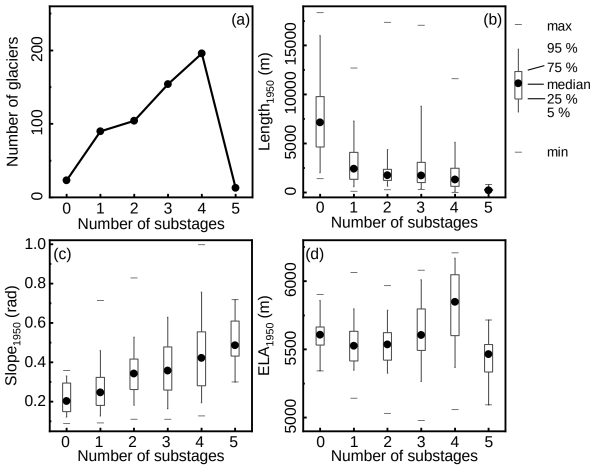

Figure 5. (a) The identified glacial substages number distribution in the MC experiment. The relationship between identified glacial substages with (b) glacier length, (c) glacier slope, and (d) glacial ELA at 1950 in the MC experiment.

The occurrence time of each glacial substage also varies from glaciers, supported by the dispersal of moraine ages (Fig. 4b). Notice that not all glaciers in BH reached their *maximum peak GLRs* during LIA-4, and taking a step back, even among the glaciers with the *maximum peak GLR* during LIA-4, the occurrence times are also different (Fig. 6a). Statistically, about 48.1 % glaciers experienced their *maximum peak GLR* during LIA-4 followed by 36.1 % glaciers reaching their *maximum peak GLR* during LIA-1. Therefore, the occurrence time of *maximum peak GLR* at regional scale is associated with the occurrence time of the majority of glaciers reaching their *maximum peak GLRs*. In addition, this can in turn explain the lack of some moraines. Considering two glaciers both having four glacial substages but different occurrence times of *maximum GLR peak* (one at LIA-4 and another at LIA-1) during LIA, we might find 4 moraines for the glacier which reaches its *maximum GLR peak* at LIA-4 but only 1 moraine for the other because the first three moraines are destroyed by the last glacier advance. Similarly, this phenomenon also remains in the occurrence times of the *second/third/fourth peak GLR* (Fig. 6b-c).

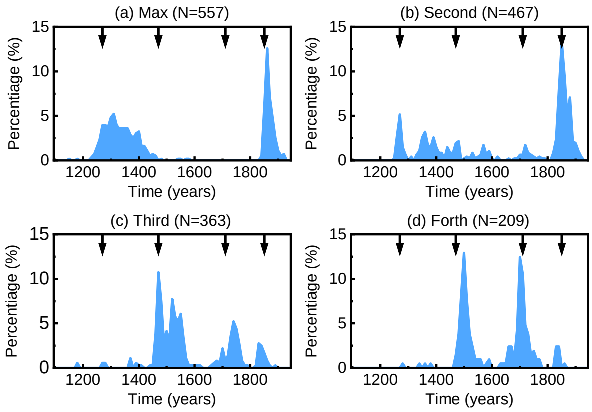

Figure 6 The percentage of the glaciers with (a) *maximum peak* GLR, (b) the *second largest peak* GLR, (c) the *third largest peak* GLR, and (d) the *fourth largest peak* GLR over time in the MC experiment. The arrows represent the time of the four glacial substages, 1270s (LIA-4), 1470s (LIA-3), 1710s (LIA-2), and 1850s (LIA-1).

In summary, four LIA glacial substages at 1270s, 1470s, 1710s, and 1850s were found in BH based on the MC experiment. The maximum glacier extent appeared during LIA-4, comparable with the moraine ages. The regional glacial evolution is a collective effect of individual glacier changes. Four substages during LIA at the regional scale does not guarantee that each individual glacier has four substages. Likewise, not all glaciers in BH reached their *maximum peak GLRs* during LIA-4. Instead, it only represents the characteristics of most typical glaciers that accounted for the vast majority of the total glaciers. This can explain why there exists four substages in regional scale in the simulation but is hard captured in the previous studies that only focus on one individual glacier, which helps us to deeply understand the relationship between regional glacial evolution and the individual glacier response to climate change.

4.3 Climate-forcing Mechanisms

Despite we have found the regional glacial evolution is a collective effect of individual glacier changes, the dominant climatic mechanisms behind are still unclear. A better understanding of the possible forcing mechanism of regional paleoglacier fluctuations at centennial timescales benefits projecting glacier outlooks in the future (Solomina et al., 2015). However, due to the limitation of field investigations, previous studies simply ascribed the glacier change to the temperature variation in the monsoon-influenced Himalaya by comparing the regional glacial sequences with the $\delta^{18}O$ record from Greenland, Tibetan or North Atlantic (Peng et al., 2019, 2020). As the model can explicitly link the glacier changes with climate forcings (PDD scheme), it provides us an opportunity to further explore on this issue.

Our study revealed that the regional glacial fluctuations are related to the temperature changes rather than the precipitation change (Fig. 4a and Fig. 7a, b, d). Four cold intervals around 1320s, 1510s, 1760s, and 1870s in the MC experiment corresponds to LIA-4 (1270s), LIA-3 (1470s), LIA-2 (1710s), and LIA-1 (1850s), respectively. However, this signal cannot be detected in precipitation changes. Results from six GCM experiments also support this argument though with different time and strength. The four cold intervals during the LIA in BH are closely linked to four large stratospheric sulfur-rich explosive eruptions events (sulfate aerosol loadings > 60 Tg; Fig. 7c; Gao et al., 2008). The beginning of oldest cold period (LIA-4) might be forced by a series of volcanic activities, including a massive tropical volcanic eruption in 1257 followed by three smaller eruptions in 1268, 1275, and 1284 (Miller et al., 2012). The volcanoes Billy Mitchell (1580), Huaynaputina (1600), Mount Parker (1641), Long Island (1660), and Laki (1783) may contributed to the cooling events during LIA-3 and LIA-2 (Jonathan, 2007). The 1815 eruption of Tambora and the 1883 eruption of Krakatau are believed to promote the youngest cold period of LIA (LIA-1; Rampino and Self, 1982).

Although temperature determines whether BH can run into a glacial substage, precipitation still has the ability to regulate the time of glacier advancing to its maximum in a glacial substage due to the fact that SMB is determined by the combination of temperature and precipitation according to the PDD scheme (Eq. 1; Marzeion et al., 2012; Maussion et al., 2019). Positive or negative SMB determines whether a glacier advances or retreats, and the amplitude of glacier change is directly influenced by the amplitude of SMB change and the duration of the positive or negative SMB (Marzeion et al., 2012; Maussion et al., 2019; Fig. 4a and Fig. 7e). Four peaks of SMB have been found in the MC experiment, around 1260s, 1460s, 1670s and 1820s, corresponding to each substage. Stronger precipitation, associated with larger SMB, at the beginning of the cold interval will drive the glacier advance rapidly, shortening the time for it to reach its maximum extent. In addition, we also found ELA has a good correlation of the SMB, which can be used as a proxy for SMB. ELA is the elevation where accumulation equals ablation for a certain glacier (Fig. 7f; Benn & Lehmkuhl, 2000; Heyman, 2014). Four periods of ELA dropping around 1270s (-132.2 m), 1470s (-115 m), 1690s (-113.4 m), and 1820s (-112 m) are detected in the MC experiment, agreeing well with SMB change. This finding, to some extent, would benefit field investigation as paleo ELA is easily available while paleo SMB is hard to measure.

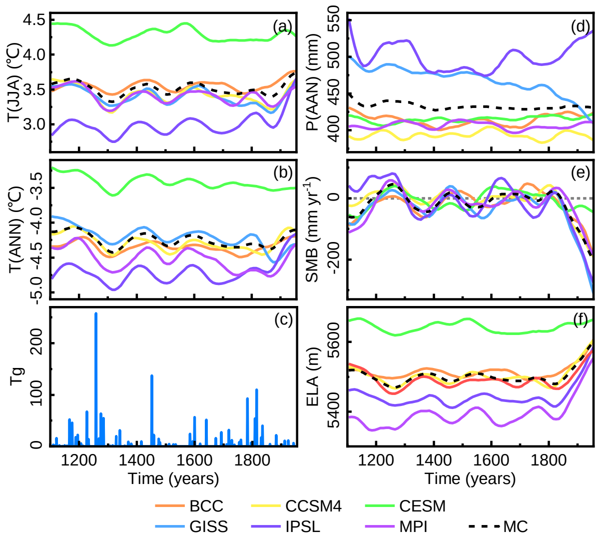

Figure 7. The regional average (a) summer temperature (T(JJA)), (b) annual temperature (T(ANN)), (d) annual precipitation (P(ANN)), (e) SMB, (f) ELA from 1100 to 1950 CE at a decadal timescale. (c) Global stratospheric sulfate aerosol loadings (Gao et al., 2008).

Seasonal climate is believed to have more important impacts on glacier evolutions than annual climate (Yan et al.,2020, 2021). We calculated the regional average monthly temperature, precipitation, cumulative SMB anomalies (relative to 1950s) from 1100s to 1950s (Fig. 8a, b, and c) in the MC experiment to investigate the effect of monthly climate changes on glacial fluctuation. Consistent with the Fig. 7e, four significant increasing periods of monthly SMB changes around 1270s, 1470s, 1710s, and 1850s are identified (Fig. 8c) as a result of monthly temperature decreasing (Fig. 8a). Monthly precipitation does not show obvious change, expect for an abrupt increasement in August (Fig. 8b). The abnormal increasing of August precipitation is polluted by the GISS climate dataset which suffers from large precipitation bias.

Strong cumulative SMB change only occurs in JJA despite the temperature change almost uniformly distributes and precipitation slightly variates (excluding August) throughout the year. The pattern of seasonal SMB change indicates that the summer temperature might dominates the annual cumulative SMB. This is because JJA is the main ablation season of glaciers in the monsoon-influenced Himalaya due to a higher temperature (Fig. 8d). A reduction of summer temperature will not only decrease the number of positive degree days but also decrease the average temperature during the positive degree days, resulting in the reduction in summer ablation (Eq. 1). Meanwhile, JJA is also the wettest season in the study area. Decreasing temperature will lead to an increasing probability of solid precipitation, enhancing the accumulation. As the SMB is determined by the sum of ablation and the accumulation, the JJA SMB is largely increased. However, though the temperature also decreases in DJF, no more precipitation will not increase the accumulation. Therefore, the SMB change is weak.

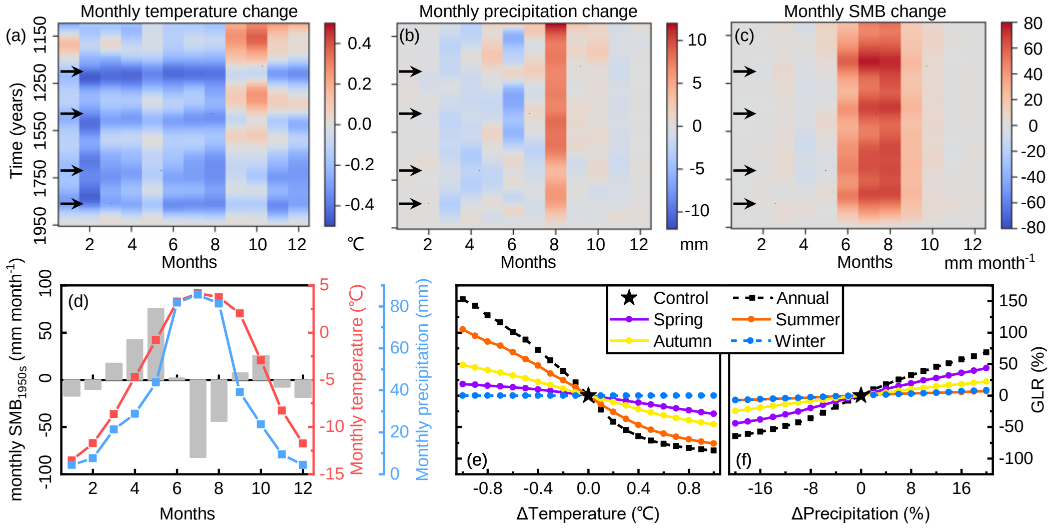

Figure 8. The monthly (a) temperature, (b) precipitation, and (c) SMB changes relative to 1950s at a decadal timescale in the MC experiment. The arrows in (a) – (c) represent the time of the four glacial substages, 1270s (LIA-4), 1470s (LIA-3), 1710s (LIA-2), and 1850s (LIA-1). (d) the monthly temperature, precipitation, and SMB distribution in 1950s. Sensitivity of GLR to annual or seasonal (e) temperature and (d) precipitation.

We also conducted a series of the sensitivity tests to examine the influence of seasonal temperature or precipitation on BH glacier change (Fig. 8e, f). Glaciers retreats gradually as a response to the temperature increases or precipitation decreases.

The sensitivity of glaciers to temperature/precipitation (changing rates) increases with temperature/precipitation increases firstly, reaches maximum at 0, and decreases after that, with an average changing rate of -160.1 %/°C for temperature and of 4.0 %/% for precipitation. Glaciers are most sensitive to summer temperature change with an average change rate of 110.4 %/°C, followed by autumn (51.6 %/°C) and spring (25.2 %/°C). Glaciers is not sensitive to winter temperature change (0.0 %/°C), supporting the results in Fig. 8c. This indicates that the temperature changes in warm seasons, especially in summer, explain the most variance of the GLR changes. Fixing temperature, the sensitivity of glaciers to precipitation changes is higher in spring (2.4 %/%), followed by autumn (1.1 %/%), summer (0.4 %/%) and winter (0.4 %/%). Therefore, the precipitation change in spring and autumn has larger influences on glacier evolution. In order to compare the relative sensitivity of temperature and precipitation to glacier change, we introduced an index $k = \frac{\Delta p}{\Delta T}$, which represents how precipitation will change in response to the temperature variation (Jeevanjee and Romps, 2018). This is an index only related to the local climate and is about 1.7 %/°C in the MC experiment. From our sensitivity tests, we need a $k = 53.0$ %/°C to maintain the LIA glacier pattern (GLR = 60.6 %), which is much larger than local climate k, indicating the temperature dominates the LIA glacial fluctuation in BH.

In summary, seasonal analysis and sensitivity tests indicate that the change in temperature, especially summer temperature, is the dominant forcing factor for glacier changes during the LIA (sub-orbital scale) in monsoonal influenced Himalaya. In contrast, the impact of precipitation change is limited. This conclusion has been drawn by Yan et al. (2020, 2021) at the orbital scales, but now can be extends to the sub-orbital scale. In addition, we also found that the temperature changes during LIA are closely related to volcanic activities (Gao et al., 2008; Miller et al., 2012).

5 Conclusions

We simulated the glacial evolution across BH during LIA using the coupled mass-balance and ice flow model, OGGM. Compared with the geomorphological maps and moraine ages, OGGM broadly captures the pattern of glacier length change, but underestimates its amplitude. The regional pattern of glacier changes is the collective effect of each glacier. The dispersal of the observations could be reproduced by the model due to the individualities of each glacier. On the regional scale, four LIA substages were identified at about 1270s, 1470s, 1710s, and 1850s (from LIA-4 to LIA-1) in the MC experiment. The most extensive glacial advances occurred during LIA-4, consistent with regional glacial chronological and geomorphic evidence. The number of glacial substages for individual glacier has a positive correlation with glacier slope. The regional glacier advances are dominated by the reduction of summer ablation.

Although limitations still exist in the simulations, such as the application of sOGGM on amplitude of glacier changes, this study presented the first simulation of sub-millennium glacial evolutions during LIA in BH using the OGGM. We found a testable relationship between seasonal climate change and glacier expansion, explained the dispersal of moraine ages and revealed the reasons for the four glacial substages during LIA in BH. Our findings link the limited observations with the model simulations and provide important insights into the climate forcing mechanism on glacier change at centennial timescale.

Code and data availability. Code to run OGGM v1.5.0 is available at https://zenodo.org/record/4765924#.YnYuB4dBxD8 (Maussion et al., 2019).

Author contributions. Study concept devised by CW. YW performed the model runs and analysis, and wrote the original draft. LY and LG reviewed and revised the paper.

Competing interests. The authors declare that they have no conflicting interests.

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