

Sensitivity of the Asian Monsoon to Little Ice Age Mountain Glacier Expansion

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Abstract

Mountain glaciers cover less than 1% of Earth's land surface, yet their influence on climate can reach far beyond the ice itself. Using the fully coupled INMCM6 Earth System Model, we quantify how expanding mountain glaciers to their Little Ice Age (LIA) extent perturbs the Asian monsoon under otherwise pre-industrial conditions. The prescribed glacier mask raises surface albedo over the Tibetan Plateau by 1.6–1.8% and cuts the sensible heat flux by about 3 W m^{-2} , cooling the overlying column and lowering mid-tropospheric geopotential heights. This contributes to the strengthening of the western subtropical jet stream and its shift to the south of the Plateau. Through the pre-monsoon season, forming an upper-level dynamical barrier that suppresses the northward advance of the rain belt. The imprint is a redistribution rather than a collapse: onset-phase (April-June) precipitation along the southern Plateau margin falls by roughly 20%, an amount equivalent to a delay in the monsoon about two to three weeks. These changes are accompanied by drying over the Tibetan-Himalayas region and modest moistening displaced toward eastern China and the Korean Peninsula. Crucially, the large-scale engine of the monsoon is left intact - neither the Somali Jet nor four standard circulation indices change significantly. LIA glacier expansion therefore retimes and spatially reorganizes monsoon rainfall through upper-level dynamics, a regional signal that bulk monsoon indices largely miss.

1 Introduction

Mountain glaciers cover less than 1% of the Earth's surface, yet they play a disproportionately important role in regional climate systems through their coupling with the atmosphere (Sauter et al. 2025). By modifying surface albedo, roughness, and the surface energy and mass balance, glaciers influence the local radiation and heat budgets, the hydrological cycle, and near-surface circulation, with effects that can extend beyond the glacierized areas themselves (Salerno et al. 2023). Understanding these glacier-climate feedbacks is therefore essential for describing cryosphere-atmosphere interactions, both under present-day warming and during past climate states.

Numerical modeling is one of the few tools that allow these interactions to be studied systematically, and it has developed along two complementary lines. On the one hand, dedicated glacier models resolve the physics of individual glaciers or glaciated regions, ranging from detailed process-based surface energy- and mass-balance models (e.g., Steiner et al. 2018; Marshall 2021) to models of intermediate complexity that trade some physical detail for computational efficiency. Among the latter, reduced-complexity approaches ranging from volume-area scaling (Bahr et al. 2015) and empirical retreat parameterizations (Huss et al. 2010) to linear response models (Roe and Baker 2014) - reproduce the first-order dynamic response of a glacier from a small number of geometric and climatic parameters, which makes them attractive as building blocks for large-scale models (Zekollari et al. 2022). Among these, the minimal glacier model of Oerlemans (2008) has proven a convenient basis for describing mountain glaciation and has been applied in this context by Toropov et al. (2023). Building on this approach, global glacier models (GGMs) such as IGRICE couple an Oerlemans-type dynamic core with an explicitly simulated surface mass balance that accounts for orographic precipitation, radiation

redistribution over the glacier surface, turbulent heat fluxes, and snow-cover evolution on ice (Toropov et al. 2025). Being both physically founded, and computationally inexpensive, such models are increasingly regarded as a natural basis for representing mountain glaciers within Earth system models.

On the other hand, Earth system and coupled climate models (ESMs) have been used to assess the climatic imprint of mountain glaciation, typically through simplified representations of glaciers or prescribed cryospheric forcing (e.g., Jin et al. 2005; Goswami et al. 2022; Cao et al. 2022). Most contemporary ESMs, however, still lack an interactive mountain-glacier component, and the incorporation of dedicated glacier parameterizations is only beginning (Toropov et al. 2025). Therefore, the sensitivity of large-scale atmospheric circulation to changes in mountain glaciation - and the extent to which ESMs are able to reproduce it - remains poorly constrained. This gap between glacier-scale process modeling and coarse-resolution ESM experiments motivates the present study.

While the near-surface effects of glaciers are comparatively well documented, their imprint on the free troposphere and on the large-scale circulation is far less certain. A sufficiently large change in glacierized area can perturb the mid-tropospheric temperature and pressure fields, and through them the position and intensity of regional circulation systems at considerable distance from the ice itself (Salerno et al. 2023). The magnitude and spatial reach of this response remain poorly quantified in global climate models.

The Little Ice Age (LIA; roughly the 14th-19th centuries) provides a natural framework for examining glacier-climate interactions during a period of widespread mountain-glacier expansion (Lüthi, 2014). At its peak, glaciers in many regions advanced well beyond their present-day limits, with terminus positions displaced by up to several kilometers relative to modern values (Lee et al. 2021). Multi-proxy reconstructions estimated, the mean annual temperature decreased about 0.5°C averaged in the North hemisphere, with a maximal negative anomaly in the North Atlantic region (about 1.5–2°C) (PAGES 2k Consortium, 2013; Schurer et al. 2017). Rather than resulting from a single cause, this cooling is now understood as the response of the coupled climate system to a combination of external forcings and internal feedback. Clustered explosive volcanism is increasingly regarded as the primary trigger: closely spaced sulfur-rich eruptions produced abrupt summer cooling, which was then sustained for decades to centuries by sea-ice and ocean-circulation feedback in the North Atlantic, long after the volcanic aerosols had been removed (Miller et al. 2012; Brönnimann et al. 2019). Reduced solar irradiance during grand solar minima, most notably the Maunder Minimum, has traditionally been invoked as an additional driver, although recent reconstructions suggest that its direct radiative contribution was comparatively modest (Owens et al. 2017; Schurer et al. 2014). Because the LIA combines these natural forcings with minimal anthropogenic influence, it is particularly well suited for isolating natural glacier-climate feedback, and the accompanying expansion of mountain glaciation can be treated as a natural experiment for studying its imprint on regional atmospheric circulation. Here we isolate the glacier-albedo component of this forcing, holding other boundary conditions at pre-industrial levels

To address this gap, we employ the INMCM6 Earth System Model (Volodin et al. 2024), which provides a fully coupled representation of atmosphere, ocean, land surface, sea ice, aerosols, and the carbon cycle. Unlike studies focusing either on individual glacier processes or on simplified cryospheric forcing, our approach incorporates a reconstructed LIA mountain glacier mask into an Earth system model to quantify the sensitivity of large-scale atmospheric circulation and the Asian monsoon to glacier expansion under pre-industrial conditions. The paper is organized as follows. Sec. 2 briefly describes the INMCM6 model, the preparation of the detailed mountain glacier mask, and setting up numerical experiments. In Sec. 3, we quantify the sensitivity of key atmospheric variables to the glacier module and characterize how far from the glacierized regions, and over which seasons, the atmospheric response develops in INMCM6. In Sec. 4, we discuss the physical mechanisms behind the diagnosed changes and their consistency with previous modeling studies and proxy-based reconstructions of LIA climate. Finally, conclusions and suggestions for the development of ESMs are provided in Sec 5.

2 Materials and methods

2.1 The model

The INMCM6 (Volodin et al. 2024; Bragina et al. 2024) is a fully coupled Earth System Model developed at the Marchuk Institute of Numerical Mathematics RAS. It comprises interactive atmosphere, ocean, sea-ice, land-surface, aerosol, and carbon-cycle components that are integrated together and exchange fluxes at every time step, so that the simulated climate emerges from their mutual interaction rather than from prescribed boundary conditions. This coupling is essential for the present study: an imposed glacier perturbation simultaneously alters the surface energy and water balance, the strength of land-atmosphere coupling, and, through the resulting heating anomalies, the large-scale circulation. A fully coupled framework is therefore required to capture not only the local response over the glaciated surface but also the remote, dynamically communicated atmospheric response that is the focus of this work.

The atmospheric component integrates the full set of primitive equations (hydrostatic momentum, thermodynamic, continuity, and moisture equations) on a latitude-longitude grid in a terrain-following σ vertical coordinate, using a finite-difference dynamical core formulated to conserve mass and energy. Sub-grid processes are represented by a standard suite of physical parameterizations: a broadband radiative-transfer scheme accounting for greenhouse gases, aerosols, and clouds; mass-flux moist (deep and shallow) convection together with a large-scale condensation and prognostic cloud scheme; a planetary-boundary-layer and turbulence scheme; orographic gravity-wave drag; and a multilayer land-surface scheme resolving soil heat and moisture, vegetation, and snow. An interactive aerosol block carries the main species (sulfate, black and organic carbon, sea salt, and dust), and an interactive carbon cycle couples terrestrial and oceanic reservoirs. The ocean component solves the primitive equations in σ -coordinates and is coupled to a dynamic-thermodynamic sea-ice model.

The INMCM model family participates in the Coupled Model Intercomparison Project: the previous version, INM-CM5-0, was an official CMIP6 contributor (Volodin et al. 2017), and the present INMCM6 version has been run following the CMIP6 protocol, including the historical experiment (1850–2014) and the Scenario MIP SSP projections (Gritsun et al. 2024), under the standard CMIP6 experimental design (Eyring et al. 2016). This lineage ensures that the model configuration used here is consistent with a well-documented and widely evaluated modelling framework.

In this study, the base configuration of INMCM6 is used. It simulates the global atmospheric circulation with a horizontal resolution of $2^\circ \times 1.5^\circ$ in longitude and latitude and 21 vertical σ -levels (the uppermost level near 30 km); the ocean general-circulation component has a horizontal resolution of $1^\circ \times 0.5^\circ$ and 40 vertical σ -levels. This coarse but well-tested configuration is well suited to long paleoclimate and single-forcing sensitivity experiments such as the present one. By default, however, INMCM6 does not represent mountain glaciation, so for this study we developed a dedicated model version that includes a simplified mountain-glacier effect (Section 2.3).

2.2 Preparation of mountain glacier mask

To investigate glacier-climate interactions during the LIA, a mountain glaciation mask is prepared. The LIA maximum was selected because it represents the most recent period of widespread glacier expansion prior to the instrumental climate record. The modern glacier distribution is taken as the baseline, derived from two inventories: the Randolph Glacier Inventory (RGI 7.0 Consortium, 2023) and the Glacier Catalogue of Russia (Institute of Geography RAS, 2025). To approximate LIA conditions, the modern glaciated areas in each mountain system are scaled upward using regional area loss estimates reported in the literature. Specifically, the scaling factors are derived from reconstructions of glacier terminus positions during maximum advances in the LIA for the following regions: the Himalayas (Lee et al. 2021), the Alps (Zasadni 2007), Scandinavia (Baumann et al. 2009), Svalbard (Martín-Moreno and Allende-Álvarez 2016), the Rocky Mountains (Luckman 2000), New Zealand (Carrivick et al. 2020), Alaska (Curran 2017), and the Caucasus (Tielidze et al. 2025). These regional reconstructions rely primarily on the geomorphological mapping of glacial landforms - terminal and lateral moraines and trimlines that delineate the LIA-maximum position of each glacier. The age of these landforms is constrained by dating techniques such as lichenometry, dendrochronology, and historical or cartographic records, while in several studies the former glacier geometry is further inferred from reconstructed equilibrium-line altitudes or ice-thickness modelling. Together they provide regionally consistent estimates of glacierized area at the LIA maximum relative to the present. Technically, the mask gives the spatial distribution of land cell fraction covered by glacier. The data are initially processed to a spatial resolution of $0.5^\circ \times 0.5^\circ$ and then interpolated to the INMCM6 atmosphere and land-surface grid ($2^\circ \times 1.5^\circ$).

We therefore assume that the omission of small glaciers that disappeared completely after the LIA introduces only a second-order uncertainty relative to the imposed regional changes in glacierized area. This limitation should be considered when interpreting local-scale responses but is unlikely to affect the large-scale circulation anomalies investigated in this study.

The reconstructions used here document substantial historical glacier losses. For example, a 40% area reduction in Himalayan glaciers since the Little Ice Age, and volume estimates of $280 \pm 43 \text{ km}^3$ for Alpine glaciers around 1850 with subsequent reductions of 64% by 2015. Scandinavia glacier area decreased by 84 km^2 from the 1960s to 2010s. Trends in glacier shrinkage are also noted in the Rocky Mountains, New Zealand (with doubled melt rates since pre-industrial times), and other regions.

The glaciers listed above reached their maximum extents asynchronously, which makes a uniform approximation of a single 19th-century climate state imprecise. Nevertheless, using their maximum areas available for this period gives a more representative depiction of the cryosphere during peak glacier expansion periods. This method avoids using a single arbitrary date that may not coincide with the actual maximum values for many glaciers. The proposed approach is consistent with contemporary glaciological practices and the LIA climate reconstructions (Reinthal and Paul 2025). Precise areal estimates for all mountain glaciers remain unavailable, necessitating manual processing of individual glaciers.

2.3 Numerical experiments description

In glacier mode for INMCM6, land surface covered by glaciers is specified by the prescribed mask. We assume that this area is covered in snow, with the corresponding albedo and other properties. According to the prepared mask (Section 2.2), the glacier fraction (`ice_land_ratio`) is given for each land grid cell. Effective snow-covered fraction (sfr_{eff}) is defined by Eq. (1):

$$sfr_{eff} = sfr + (1.0 - sfr) \times \text{ice_land_ratio}$$

1

Here, the value `sfr` is a snow-covered fraction for corresponding land grid cell without accounting of glaciers (as in the base version of INMCM6). It depends on snow water equivalent and vegetation type (e.g., low grass is fully masked by snow, while forest canopy is only partially masked). Snow-covered fraction determines the average grid cell albedo as well as the land surface heat balance.

The described approach raises the effective snow-covered fraction over glaciated cells independently of the simulated snowpack, while the standard vegetation-dependent snow masking is retained elsewhere.

In this research, we carry out a series of pre-industrial model experiments simulated the Earth climate around 1850. Throughout the simulation all external model forcings (concentration of greenhouse gases, natural and anthropogenic emissions, aerosols, Solar constant and Solar spectrum, and land use) are fixed at levels corresponding to 1850, that is considered an appropriate approximation of the pre-industrial climate state (Eyring et al. 2016). For this purpose, CMIP6 forcing dataset was used. The pre-industrial experiment simulates conditions representative of the late LIA. Moreover, this setup captures the internal model dynamics and allows to evaluate clear feedback on the only one forcing (mountain glaciers in this research) holding the others fixed.

The simulation duration for all experiments is 75 years. The first 25 years serve as a spin-up phase to achieve model equilibrium; the analysis is based on the subsequent 50 years. All differences between the glacier and control experiments are therefore evaluated on 50 annual values per experiment, which sets the sample size for the significance testing described in Sect. 2.3.

Simulations can be divided into two parts: the control experiment with the default INMCM6 model and the experiment using a specially prepared model version that includes the effects of mountain glaciers. It uses the mask incorporating glaciers at their maximum extent during the LIA (Fig. 1a). Relative to the present-day glacier distribution, the prescribed LIA mask increases the glacierized area by $30.3 \times 10^3 \text{ km}^2$ (+ 15.2%) over North America and Greenland and by $41.3 \times 10^3 \text{ km}^2$ (+ 37.0%) over Northern Eurasia and High Mountain Asia (Fig. 1b, c). The expansion is realized almost entirely as an increase of the cover fraction within grid cells that are already glacierized today: cells carrying LIA ice but no present-day ice contribute only 1.7% and 6.5% of the respective area increments (3 and 8 grid cells). Local differences in cover fraction reach 8% of the grid-cell area, with the largest values over the Alaska-Yukon ranges, the Canadian Arctic Archipelago, the Greenland margins, Svalbard, Novaya Zemlya and the Himalaya-Tibetan Plateau sector. Because the mask is defined on the $1.5^\circ \times 2.0^\circ$ model grid, these figures describe the glacierized fraction resolved by the model and are not directly comparable with glacier inventories at native resolution.

The simulation duration for all experiments is 75 years. The first 25 years serve as a spin-up phase to achieve model equilibrium. For the analysis of results, we focus on the next 50 years..

2.4 Statistical analysis

Throughout this study, an anomaly is defined as the difference between the LIA and the control (piControl) experiment, LIA minus Control, computed at each grid point and for each season from the 50-year climatological means of the two runs (model years 26–75):

$$\Delta_i = \bar{x}_{LIA,i} - \bar{x}_{CTRL,i}$$

2

where $\bar{x}_{LIA,i}$ and $\bar{x}_{CTRL,i}$ denote the 50-year seasonal mean of a given variable at the i -th grid point in the LIA and Control experiments, respectively. Positive values therefore indicate larger values in the LIA experiment. No observational data are used as a reference: all quantities reported below characterize the model response to the imposed glacier forcing, not model error relative to observations.

Statistical significance was assessed at each grid point with a two-sided Welch's t-test for unequal variances (Welch 1947), applied to the two seasonal time series of 50 annual values each ($n_{LIA} = n_{CTRL} = 50$), constructed from model years 26–75 to exclude the initial adjustment phase. The test statistic is

$$t = \frac{\bar{x}_{\text{LIA}} - \bar{x}_{\text{CTRL}}}{\sqrt{s_{\text{LIA}}^2/n_{\text{LIA}} + s_{\text{CTRL}}^2/n_{\text{CTRL}}}}$$

3

where s^2 is the sample variance and the degrees of freedom (which vary from point to point) follow the Welch-Satterthwaite approximation. The null hypothesis of equal means was rejected at the 90% confidence level ($p < 0.1$); this threshold defines the significance mask used in all figures. Significance is applied descriptively, to delineate the responding regions; no multiple-testing (field-significance) correction was applied.

To characterize the amplitude and spatial structure of the response, we additionally report two integral metrics -the mean anomaly (Bias) and its root-mean-square (RMSE):

$$\text{Bias} = \frac{1}{N} \sum_{i=1}^N \Delta_i$$

4

$$\text{RMSE} = \sqrt{\frac{1}{N} \sum_{i=1}^N \Delta_i^2}$$

5

where the sum runs over the N grid points of the analysis domain (50–150°E, 15°S–50°N) at which the anomaly is statistically significant ($p < 0.1$). These are not measures of model skill against observations but integral measures of the response over the responding area; because the sample is restricted to significant points, they are systematically larger in magnitude than the corresponding full-domain averages. The ratio $|\text{Bias}|/\text{RMSE} \in [0, 1]$ indicates the degree of spatial cancellation between positive and negative anomalies, with values near 1 denoting a spatially coherent, single-signed response and values near 0 a strongly compensating one.

Finally, Tables 1–2 report an area-integrated significance statistic t_A , distinct from the grid-point statistic of Eq. (3): t_A is a Welch two-sample statistic (unequal variances) comparing the cos-latitude-weighted area-mean seasonal time series of the two experiments over the significant grid points. Whereas the grid-point test of Eq. (3) is applied independently at each point to build the significance mask, Bias, RMSE and t_A are domain-integrated diagnostics. The statistical procedures follow standard practice in the atmospheric and climate sciences (Wilks 2011).

3 Results

3.1 Regional energy balance response

The extensive glacier coverage in the LIA experiment notably changes the surface energy budget over the Himalayas and Tibetan Plateau. These changes serve as the primary thermal forcing for the regional climate anomalies discussed in subsequent sections. Throughout the analysis we use the two standard monsoon seasons: the pre-monsoon season MAM (March-May) and the summer monsoon season JJAS (June-September). These definitions follow common practice in South Asian monsoon research and allow direct comparison with previous studies.

Table 1

Basic statistics of the significant ($p < 0.1$) anomalies for MAM, spatially averaged over the entire map domain of Fig. 2 (50°-150°E, 15°S-50°N).

Mean	Albedo (%)	SW balance (W/m ²)	LW balance (W/m ²)	Net balance (W/m ²)	Sensible heat (W/m ²)	Latent heat (W/m ²)
	1.8	-2.5	2.0	-1.5	-2.8	0.1
Median	1.8	-3.6	2.0	-2.2	-2.6	-0.1
Std	3.1	4.3	2.3	2.8	3.1	4.2
Min	0	-9.7	-3.8	-7.6	-12.3	-5.7
Max	4.0	7.1	6.8	4.8	1.7	12.1
Bias	1.8	-2.5	2.0	-1.5	-2.8	-0.1
RMSE	2.0	5.0	3.1	3.2	4.1	4.4
t	5.2	-3.4	4.1	-3.3	-5.5	2.4

Table 2

Basic statistics significant ($p < 0.1$) anomalies for JJAS spatially averaged over the entire map domain of Fig. 2 (50°-150°E, 15°S-50°N).

Mean	Albedo (%)	SW balance (W/m ²)	LW balance (W/m ²)	Net balance (W/m ²)	Sensible heat (W/m ²)	Latent heat (W/m ²)
	1.6	-0.9	1.7	0.1	-3	0.1
Median	1.1	-2.3	2.5	1.1	-2.5	1.6
Std	1.5	4.0	2.0	3.13	3.1	4.7
Min	0	-11.3	-4.0	-10.7	-11.5	-21.9
Max	7.6	6.6	4.4	5.7	6.0	5.5
Bias	1.6	-0.9	1.7	0.1	-3	-0.1
RMSE	2.0	4.1	2.6	3.1	4.3	4.7
t	31.2	-2.3	4.2	0.4	-11.4	0.1

The direct impact of glaciation is most evident in surface albedo (Fig. 2a, b). The anomalies are robust and highly significant, with spatially averaged increases of 1.6% (JJAS) and 1.8% (MAM) across the domain (Tables 1–2). T-statistics for albedo are exceptionally high (31.2 for JJAS), confirming the signal's reliability. Locally, maximum anomalies reach + 7.6%, particularly in central glaciated valleys. The high standard deviations (1.5–3.1%) reflect strong spatial heterogeneity, consistent with the complex topography and differential melting patterns observed in remote sensing studies (Ming et al.2015). This albedo increase drives a reduction in the Net Shortwave radiation balance (mean deficit of $\sim 2.5 \text{ W/m}^2$ in MAM), reducing the energy available for surface heating.

The reduction in absorbed solar radiation leads to a significant decrease in the Sensible Heat Flux from the surface to the atmosphere. The domain-averaged anomaly is consistently negative: -3.0 W/m^2 (JJAS) and -2.8 W/m^2 (MAM). The high absolute values of t-scores (11.4 and 5.5, respectively) indicate a coherent cooling signal across the entire glaciated region. This reduction in sensible heating is the primary mechanism for the tropospheric cooling described later in Section 3.3.

In contrast, Latent Heat Flux anomalies show a near-zero mean ($\sim 0.1 \text{ W/m}^2$) but very high standard deviation ($\sim 4.5 \text{ W/m}^2$). This indicates a patchy response: evaporation is suppressed over cold glacier surfaces but may be enhanced in adjacent areas due to circulation changes. The high variability aligns with hydroclimatic complexity described in High Mountain Asia studies (Steiner et al.2018).

The combined effect of these fluxes results in a negative total heat balance during the critical pre-monsoon season (MAM mean: -1.5 W/m^2). This energy deficit delays the seasonal warming of the Tibetan Plateau, weakening the land-sea thermal contrast that drives the monsoon onset. In JJAS, the signal becomes mixed (mean $\sim 0.1 \text{ W/m}^2$), suggesting that once the monsoon is established, atmospheric feedback (e.g., cloud cover) complicate the surface budget.

3.2 Regional Atmospheric Circulation Response

The heat balance anomalies described in the previous section govern the cooling of the troposphere over the Tibetan-Himalayan region, which may influence monsoon circulation. The glacier area anomaly in a LIA numerical experiment initiates a dynamical chain from the surface heat balance to the large-scale regional circulation. The primary atmospheric response to the surface cooling should be the reduction in the thickness of the atmospheric column over the Tibetan Plateau. Figure 3 (a, c) illustrates the anomalies of geopotential height at 500 hPa and a corresponding response in mid-tropospheric winds.

In the pre-monsoon season (Fig. 3a), a statistically significant depression of the geopotential height is evident, centered directly over the Plateau and extending into Central Asia. This negative anomaly reflects the reduction in tropospheric thickness caused by the cooling of the middle troposphere. Vertical temperature profiles confirm that this cooling signal propagates upward to approximately 400 hPa, significantly sharpening the meridional temperature gradient between the cold Plateau and the warm tropical Indian Ocean. In the monsoon season (JJAS, Fig. 3c), the cyclonic anomaly persists but

weakens and shifts northward, consistent with the reduced land-sea thermal contrast during the summer.

The sharpening of the meridional temperature gradient induces a robust response in the upper-tropospheric winds via the thermal wind relation. Figure 3 (b, d) presents the anomalies of the zonal wind component at 200 hPa, revealing the critical dynamical mechanism behind the delayed monsoon. During MAM (Fig. 3b), a striking dipole pattern emerges a coherent band of accelerated westerly winds (anomalies $> +1.9$ m/s) stretches across Northern India and the southern flank of the Tibetan Plateau. Concurrently, a reduction in westerly wind speed is observed further north (40°N) over Central Asia, confirming the equatorward displacement of the jet stream axis.

This dipole signifies a southward shift and intensification of the Subtropical Westerly Jet (STJ). Climatologically, the northward migration of the STJ off the Tibetan Plateau is a prerequisite for the South Asian monsoon onset (Schiemann et al. 2009; Zhang et al. 2019). Our simulation demonstrates that the glacial cooling effectively anchors the jet south of the Plateau during the pre-monsoon season. This reinforced dynamical barrier inhibits the northward propagation of the monsoon trough and suppresses convective instability, providing the physical explanation for the severe MAM precipitation deficits identified in Section 3.3.

During JJAS (Fig. 3d), the anomaly field represents a change in the wind vectors themselves, with an anomalous westerly (positive zonal) component of ~ 1.0 m/s over the Iranian Plateau and Northern Pakistan. This indicates that the "glacial brake" on the circulation continues to operate through the summer, likely contributing to the weakening of the meridional flow in the lower troposphere at 850 hPa (not shown) and to the reduced moisture transport into the continental interior.

3.3 Changes in the moisture regime

The incorporation of LIA mountain glaciers triggers a noticeable restructuring of the Asian monsoon hydroclimate, with a distinct seasonality. During the established monsoon (JJAS) the response is a modest, spatially confined drying over the elevated Tibetan Plateau, without a coherent compensating dipole. The dominant signal instead emerges in the pre-onset window (AMJ), as an intense, spatially focused suppression of precipitation along the climatological leading edge of the advancing rain belt, which delays and weakens the monsoon onset transition.

During the established monsoon (JJAS; Fig. 4b) the glacier forcing produces a modest, spatially constrained drying over the central Tibetan Plateau and the Himalaya, with local deficits reaching -1.75 mm/day (of order -210 mm over the June-September season). The reduction is confined to elevated terrain, while surrounding low-elevation regions show weaker, mostly statistically insignificant anomalies. Isolated positive anomalies over the Indo-Gangetic Plain and the southeastern Bay of Bengal are neither spatially coherent nor uniformly significant, precluding a straightforward interpretation as a compensating dipole. The dominant JJAS signal is thus a suppression of orographic convection over the Plateau rather than a large-scale redirection of monsoon moisture.

The strongest and most timing-relevant response appears in the pre-onset window (AMJ; Fig. 4a). Diffuse negative anomalies of -0.4 to -0.8 mm/day cover the Plateau and the adjacent ranges, but they intensify into a sharply localized core at the climatological leading edge of the advancing rain belt ($\sim 88-90^{\circ}\text{E}$, $25-28^{\circ}\text{N}$), where the deficit reaches -3.96 mm/day -deeper than anywhere in the JJAS field and coincident with the $87-90^{\circ}\text{E}$, $25-30^{\circ}\text{N}$ box analyzed below. This pattern is consistent with a southward shift of the subtropical jet stream (STJ) driven by the cooling of the Tibetan Plateau - a result of a negative heat balance anomaly caused by an increase in glaciated area.

A weak positive signal over northern China and the Korean Peninsula may reflect a northeastward shift of the westerly moisture corridor under the displaced jet, though its statistical robustness is limited. The clearest diagnostic of the glacier impact on monsoon timing is the annual cycle of precipitation averaged over the $87-90^{\circ}\text{E}$, $25-30^{\circ}\text{N}$ domain (Fig. 4c) -the box that contains the AMJ deficit core and lies at the climatological leading edge of the northward-advancing rain belt, where the onset signal emerges earliest and is most sensitive to a shift in timing. In the control run precipitation rises steeply through May and June to a June maximum (~ 19 mm/day). The LIA run reproduces the same seasonal shape and a comparable summer maximum (~ 18 mm/day), but its rising limb is systematically suppressed -most strongly in May (-30%) -and its seasonal maximum is reached about two weeks later (early July rather than June). Onset-phase precipitation (April-June) is reduced by $\sim 20\%$, delaying the sharp pre-monsoon increase by roughly two weeks. By July the two curves converge, indicating that the glacier forcing perturbs the timing of onset rather than the peak intensity of the established circulation.

The month-by-month decomposition of this onset-phase deficit is shown in Fig. 4d. Precipitation is reduced in the LIA experiment throughout all three pre-monsoon months, but the suppression peaks sharply in May: April declines from 6.5 to 5.4 mm/day (-17%), May drops from 12.0 to 8.3 mm/day (-30%), and June falls from 19.4 to 16.6 mm/day (-15%). This May minimum is diagnostic: it is precisely the month in which the control climatology undergoes its steepest pre-monsoon increase. The glacier forcing therefore acts most effectively by flattening the rising limb of the annual cycle, rather than by lowering the already-established June rainfall. This confirms that the $\sim 20\%$ AMJ reduction reflects a delayed and weakened onset transition rather than a uniform seasonal drying.

Figure 5. Atmospheric response to LIA glaciation (LIA minus Control). Specific humidity anomalies (g/kg, shading) at 850 hPa overlaid with wind vector anomalies at 850 hPa as well (1 m/s reference arrow) for (a) pre-monsoon (MAM) and (b) monsoon (JJAS) seasons. Only statistically significant ($p < 0.1$) anomalies are shown.

The spatial restructuring of moisture is illustrated by the specific humidity and wind anomalies at 850 hPa (Fig. 5). During the pre-monsoon season (MAM, Fig. 5a), the dominant drying anomaly (-0.4 to -0.7 g/kg) is concentrated over the Indochina Peninsula and the adjacent maritime continent ($10-20^{\circ}\text{N}$, $90-110^{\circ}\text{E}$), with a secondary negative center over central India ($\sim 20^{\circ}\text{N}$, 80°E). In contrast to the more commonly discussed Arabian Sea pathway, the anomalies over the Arabian Sea and the Bay of Bengal

proper are weak and largely statistically insignificant in this season. The accompanying wind anomalies show anomalous northerly and northeasterly flow over the Indian subcontinent and the eastern Indian Ocean sector, consistent with the southward anchoring of the STJ described in Section 3.2. This circulation adjustment suppresses the early-season moisture advection from the south and inhibits pre-onset convective development over the Indian interior. A pronounced moistening signal ($+0.3$ to $+0.7$ g/kg) appears over central and eastern China ($25\text{--}35^\circ\text{N}$, $100\text{--}125^\circ\text{E}$), suggesting that the glacier-induced dynamical reorganization redirects moisture northeastward along the East Asian coast rather than allowing it to penetrate the Indian interior. This remote response indicates that the glacier forcing produces a large-scale redistribution of moisture across the Asian sector, with drying concentrated in the monsoon source regions and compensating moistening displaced toward East Asia.

During the monsoon season (JJAS, Fig. 5b), the pattern undergoes a notable shift. The strongest drying signal (-0.2 to -0.6 g/kg) moves to the Tibetan Plateau and northwestern China ($30\text{--}40^\circ\text{N}$, $70\text{--}90^\circ\text{E}$), consistent with the persistent surface energy deficit over glaciated terrain. A compensating moistening anomaly ($+0.2$ to $+0.6$ g/kg) develops over northeastern China and the Korean Peninsula ($30\text{--}45^\circ\text{N}$, $110\text{--}135^\circ\text{E}$), pointing to a northeastward displacement of moisture along the East Asian margin. Critically, the Bay of Bengal and northeastern India show no statistically significant humidity anomaly in JJAS, which is inconsistent with a simple monsoon weakening or dipole interpretation. The 850 hPa wind anomalies in JJAS are largely statistically insignificant across the domain, confirming that the large-scale low-level monsoon flow -including the Somali Jet -maintains its climatological strength and direction. The moisture redistribution in JJAS therefore reflects a thermodynamic reorganization driven by the upper-level dynamical barrier rather than a mechanical deflection of the low-level monsoon circulation.

4 Discussion

Our experiments demonstrate that expanding mountain glaciers to their LIA extent reorganizes the Asian monsoon system. This reorganization occurs primarily through dynamical displacement, rather than a basin-wide weakening of monsoon circulation. This distinction is crucial because a number of commonly used bulk diagnostics can remain nearly unchanged, while regional hydroclimate impacts become large and spatially heterogeneous. The physical sequence underlying this glacier-driven reorganization is summarized schematically in Fig. 6.

Our results offer a mechanistic, circulation-based reinterpretation of the classic Blanford hypothesis (Blanford 1884). Traditionally, the inverse relationship between Himalayan snow cover and monsoon intensity has been attributed primarily to the surface thermal contrast where excessive snow albedo and heat required for melting cooling reduce the land-sea thermal gradient required to drive the monsoon. However, the core pathway identified in INMCM6 experiment results extends beyond this thermodynamic view, highlighting an upper-level dynamical control lever. The expansion of glaciers triggers a surface energy deficit through higher albedo and lower sensible heat flux, as summarized in Fig. 6. As this cooling propagates upward, it causes thermal contraction of the atmospheric column, leading to lower

geopotential heights in the mid-troposphere. This thermal adjustment sharpens the meridional temperature gradient, effectively anchoring and intensifying the subtropical westerly jet (STJ) over the Tibetan Plateau during the pre-monsoon season. By mechanically delaying the northward migration of the STJ, this dynamical barrier regulates the timing of the monsoon transition more directly than surface temperature anomalies alone. The southward anchoring of the STJ creates upper-tropospheric divergence and subsidence over the Indian subcontinent, mechanically suppressing the development of deep convection required for monsoon onset a mechanism well-established in observational and modeling studies (Schiemann et al. 2009; Zhang et al. 2019). The climatological northward shift of the STJ off the Tibetan Plateau serves as a prerequisite for South Asian monsoon onset (Yeh 1958; Tao 1958; Kong et al. 2017), and our results demonstrate that glacial cooling effectively delays this critical transition.

The proposed mechanism of glacier-driven circulation reorganization aligns with earlier modeling studies that identify the Tibetan Plateau as a primary thermal driver of the Asian monsoon. Our findings with the INMCM6 model add a critical detail: the expanded glacier mask acts as a persistent forcing mechanism that anchors the subtropical jet's position during the pre-monsoon season (MAM) and early summer (JJAS) in Fig. 6. These results are consistent with process-focused studies, such as Goswami et al. (2022), which showed that Tibetan glaciation modifies surface fluxes and vertical motion. However, INMCM6 advances this understanding by reproducing these responses within a fully coupled framework that explicitly accounts for seasonal transitions.

Table 3

Definitions and simulated JJAS response of the four monsoon circulation indices. WYI - Webster & Yang (1992); SAMI - Goswami et al. (1999); SEAMI and EASMI follow the low-level zonal-wind formulation of Wang & Fan (1999). All values in m s^{-1} . Welch's t-test on the 50 annual index values of each experiment ($df \approx 94-97$); significance level 10%.

Index	Definition	Control mean (JJAS)	Delta JJAS	p (JJAS)
WYI	U850 - U200, 0-20N, 40-110E	22.55	+ 0.03	0.90
SAMI	V850 - V200, 10-30N, 70-110E	2.71	-0.13	0.18
SEAMI	U850(5-15N,90-130E) - U850(22.5-32.5N,110-140E)	3.30	+ 0.09	0.72
EASMI	U850(5-15N,90-130E) - U850(20-30N,110-140E)	4.25	+ 0.08	0.72
1				

A further notable outcome concerns four widely used large-scale monsoon circulation indices (Table 3). All four are defined for the boreal summer monsoon, either as the vertical wind shear between 850 and 200 hPa (WYI, SAMI) or as a horizontal contrast of the 850 hPa zonal wind between two boxes (SEAMI, EASMI). We therefore evaluate them for JJAS only: in the control simulation their MAM climatology are negative for SAMI, SEAMI and EASMI (-1.5 , -5.8 and -4.7 m s^{-1}) and not significantly different from

zero for WYI ($1.5 \pm 2.1 \text{ m s}^{-1}$), reflecting the fact that the monsoon shear has not yet reversed in the pre-monsoon season. Anomalies of these indices outside their design season cannot be interpreted as changes in monsoon strength.

Computed from the JJAS seasonal-mean winds of both experiments, all four indices show very small LIA–Control differences: $+0.03$, -0.13 , $+0.09$ and $+0.08 \text{ m s}^{-1}$ for WYI, SAMI, SEAMI and EASMI respectively, i.e. 0.1–5% of their climatological values, and none is significant at the 10% level ($p = 0.90, 0.18, 0.72, 0.72$). Note also that SEAMI and EASMI share their southern averaging box and differ only in the latitude band of the northern one, so they do not constitute fully independent diagnostics. Although these indices represent different sectors of the Asian monsoon, all are based on spatially averaged wind fields; they therefore describe the bulk strength of the monsoon circulation and are comparatively insensitive to the regional redistribution of circulation and precipitation documented in Section 3.3.

To verify the stability of the lower-tropospheric monsoon circulation, we examined the response of the Somali Jet (also known as the Findlater Jet (Findlater 1969)), which serves as the primary conduit for moisture transport into the Indian subcontinent. Although the INMCM6 model correctly reproduces the climatological structure of this jet, the LIA glacier experiments show no statistically significant anomalies in wind speed at 850 hPa within its core. This indicates that the large-scale moisture transport remains robust under LIA conditions, consistent with the established understanding that the jet's intensity is primarily maintained by large-scale pressure gradients and orographic features (Krishnamurti et al. 1976). The absence of a weakening signal in the Somali Jet, combined with negligible changes in low-level vorticity over the northern Arabian Sea, reinforces our finding that glacier-induced cooling does not collapse the monsoon circulation's mechanical intensity, aligning with studies that show jet speed alone is not the sole determinant of monsoonal precipitation variability (Pushpanjali et al. 2013). Instead, the results point to a reorganization of the monsoon rather than a weakening of it: the moisture-bearing monsoonal winds retain their climatological strength, but the resulting precipitation is redistributed under the influence of an upper-level dynamical constraint - a southward-shifted and more persistent Subtropical Westerly Jet. (Fig. 6). This mechanism, whereby the latitudinal position of the subtropical jet determines the northern limit of monsoon moisture penetration, has been well documented in observational and modeling studies (Schiemann et al. 2009; Sampe & Xie 2010), and redirects the moisture flux, leading to the regional precipitation redistribution observed in our simulations.

The INMCM6 response is qualitatively consistent with earlier paleoclimate sensitivity experiments showing that snow and ice anomalies over the Tibetan Plateau weaken or redistribute monsoon rainfall through altered land-ocean thermal contrast and large-scale circulation adjustments. Jin et al. (2005) demonstrated, in a coupled model of intermediate complexity, that Tibetan cryosphere forcing during the Holocene can project beyond the Plateau into remote monsoon regions via atmospheric dynamics. The present results extend this picture by resolving the seasonal sequence more explicitly: the strongest response is observed in MAM and June, precisely the window when the climatological jet transition

occurs, and a glacier-induced colling most effectively delays monsoon onset and slow down the northward transport of moisture.

The main limitation of this study lies in the highly simplified representation of mountain glaciers in INMCM6. Instead of an interactive ice-dynamics scheme, glaciers are prescribed as static mask-based surface boundary conditions (albedo, roughness, thermal properties), so processes such as glacier flow, mass-balance feedback, and meltwater-hydrology interactions are not explicitly resolved. Another limitation is related to the relatively low spatial resolution of the model used. Due to this, the used glacier mask is too coarse, that can lead to an underestimation of the impact of accounting for mountain ice.

5 Conclusions

Using the INMCM6 Earth System Model, we have isolated the atmospheric fingerprint of expanding mountain glaciers to their Little Ice Age (LIA) extent under otherwise pre-industrial conditions. The key idea: this cryospheric anomaly transforms the Asian monsoon through an upper-level dynamical pathway rather than through a region-wide thermodynamic weakening of the circulation distinction with real consequences for how paleo-monsoon variability is diagnosed.

The mechanism is a coherent, seasonally phased chain. The prescribed glacier mask raises surface albedo over the Tibetan Plateau (+ 1.6–1.8%) and suppresses the sensible heat flux (-3.0 W m^{-2}), reducing the surface heat balance through the pre-monsoon season. This deficit propagates upward, cooling the mid-troposphere and depressing geopotential heights, which sharpens the meridional temperature gradient and anchors the subtropical westerly jet (STJ) south of the Plateau. The displaced jet acts as a dynamical barrier that holds back the climatological northward jet transition, suppresses pre-onset convection, and postpones the pre-monsoon rainfall increase by about two weeks -reducing onset-phase (April-June) precipitation by roughly 20%.

Crucially, this is a redistribution, not a collapse. The most robust precipitation response is a suppression of orographic convection over the Plateau and Himalayas, accompanied by modest moistening displaced toward eastern China and the Korean Peninsula. Meanwhile the low-level monsoon engine shows no detectable weakening during JJAS: neither the Somali Jet nor any of the four standard monsoon indices (Webster-Yang, South Asian, Southeast Asian and East Asian) departs significantly from its control value, the index anomalies amounting to 0.1-5% of their climatological magnitudes. Because these bulk metrics remain within their internal variability while regional hydroclimate is substantially rearranged, paleoclimate reconstructions that lean on integrated monsoon indices alone risk underestimating the regional hydroclimatic footprint of cryosphere forcing. These conclusions come with caveats. Glaciers are represented as a static, mask-based surface boundary condition rather than an interactive ice-dynamics scheme, and the coarse model resolution smooths the steep Himalayan orography - both of which likely dampen the true local response. Even so, the experiment shows that a change confined to less than 1% of the land surface can retime the Asian monsoon: mountain glaciers

punch well above their geographic weight, which is a strong argument for equipping Earth system models with realistic glacier parameterizations for both paleoclimate and future-climate applications.

Declarations

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Data Availability Statement.

The results of simulations with INMCM6 as well as used mountain glacier mask are available upon request from the corresponding author.

Competing Interests

The authors declare no conflicts of interest.

Author Contribution

Author Contributions Statement. I.M. Sushintsev: Data curation, Formal analysis, Visualization, Writing – original draft, Writing – review & editing. P.A. Toropov: Conceptualization, Methodology, Writing – review & editing. A.Yu. Chernenkov: Investigation, Data curation, Software, Writing – review & editing. E.M. Volodin: Investigation, Methodology, Writing – review & editing. All authors reviewed and approved the final manuscript.

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References

1. Bahr DB, Pfeffer WT, Kaser G (2015) A review of volume-area scaling of glaciers. *Rev Geophys* 53(1):95–140
2. Baumann S, Winkler S, Andreassen LM (2009) Mapping glaciers in Jotunheimen, South Norway, during the ‘Little Ice Age’ maximum. *Cryosphere* 3(2):231–243. <https://doi.org/10.5194/tc-3-231-2009>
3. Blanford HF (1884) On the connexion of the Himalaya snowfall with dry winds and seasons of drought in India. *Proc R Soc Lond* 37:3–22. <https://doi.org/10.1098/rspl.1884.0003>

4. Bragina V, Volodin E, Chernenkov A et al (2024) Simulation of climate changes in Northern Eurasia by two versions of the INM RAS Earth system model. *Clim Dyn* 62:7783–7797. <https://doi.org/10.1007/s00382-024-07306-y>
5. Brönnimann S, Franke J, Nussbaumer SU et al (2019) Last phase of the Little Ice Age forced by volcanic eruptions. *Nat Geosci* 12(8):650–656. <https://doi.org/10.1038/s41561-019-0402-y>
6. Cao J, Wang B, Ma L (2019) Attribution of global monsoon response to the Last Glacial Maximum forcings. *J Clim* 32(19):6589–6605. <https://doi.org/10.1175/JCLI-D-18-0871.1>
7. Carrivick JL, Davies BJ, James WHM et al (2020) Ice thickness and volume changes across the Southern Alps, New Zealand, from the Little Ice Age to present. *Sci Rep* 10:13392. <https://doi.org/10.1038/s41598-020-70276-8>
8. Curran JH (2017) Little Ice Age maximum glacier extent, Exit and Paradise Glaciers, Alaska: US Geological Survey data release. <https://doi.org/10.5066/F75T3HJZ>
9. Eyring V, Bony S, Meehl GA et al (2016) Overview of the Coupled Model Intercomparison Project Phase 6 (CMIP6) experimental design and organization. *Geosci Model Dev* 9(5):1937–1958. <https://doi.org/10.5194/gmd-9-1937-2016>
10. Findlater J (1969) A major low-level air current near the Indian Ocean during the northern summer. *Q J R Meteorol Soc* 95(404):362–380. <https://doi.org/10.1002/qj.49709540409>
11. Goswami BN, Krishnamurthy V, Annamalai H (1999) A broad-scale circulation index for the interannual variability of the Indian summer monsoon. *Q J R Meteorol Soc* 125(554):611–633. <https://doi.org/10.1002/qj.49712555412>
12. Goswami BB, An S-I, Murtugudde R (2022) Role of the Tibetan Plateau glaciers in the Asian summer monsoon. *Clim Change* 173:29. <https://doi.org/10.1007/s10584-022-03426-8>
13. Gritsun AS, Volodin EM, Bragina VS, Tarasevich MA (2024) Simulation of modern and future climate by INM-CM6M. *Russ J Numer Anal Math Model* 39(6):329–341. <https://doi.org/10.1515/rnam-2024-0028>
14. Huss M, Hock R (2015) A new model for global glacier change and sea-level rise. *Front Earth Sci* 3:54
15. Huss M, Juvet G, Farinotti D, Bauder A (2010) Future high-mountain hydrology: a new parameterization of glacier retreat. *Hydrol Earth Syst Sci* 14:815–829
16. Institute of Geography RAS (2025) Glacier Catalogue of Russia. <https://www.glacru.ru/>. Accessed 2025
17. Jin L, Ganopolski A, Chen F, Claussen M, Wang H (2005) Impacts of snow and glaciers over the Tibetan Plateau on Holocene climate change: sensitivity experiments with a coupled model of intermediate complexity. *Geophys Res Lett* 32:L17709. <https://doi.org/10.1029/2005GL023202>
18. Kong W, Swenson LM, Chiang JCH (2017) Seasonal transitions and the westerly jet in the Holocene East Asian summer monsoon. *J Clim* 30(9):3343–3365. <https://doi.org/10.1175/JCLI-D-16-0087.1>

19. Krishnamurti TN, Molinari J, Pan HL (1976) Numerical simulation of the Somali Jet. *J Atmos Sci* 33(12):2350–2362. [https://doi.org/10.1175/1520-0469\(1976\)033%3C2350:NSOTSJ%3E2.0.CO;2](https://doi.org/10.1175/1520-0469(1976)033%3C2350:NSOTSJ%3E2.0.CO;2)
20. Lee E, Carrivick JL, Quincey DJ et al (2021) Accelerated mass loss of Himalayan glaciers since the Little Ice Age. *Sci Rep* 11:24284. <https://doi.org/10.1038/s41598-021-03805-8>
21. Li J, Zeng Q (2002) A unified monsoon index. *Geophys Res Lett* 29(8):1274. <https://doi.org/10.1029/2001GL013874>
22. Luckman BH (2000) The Little Ice Age in the Canadian Rockies. *Geomorphology* 32, no. 3–4: 357–384. [https://doi.org/10.1016/S0169-555X\(99\)00104-X](https://doi.org/10.1016/S0169-555X(99)00104-X)
23. Lüthi MP (2014) Little Ice Age climate reconstruction from ensemble reanalysis of Alpine glacier fluctuations. *Cryosphere* 8:639–650. <https://doi.org/10.5194/tc-8-639-2014>
24. Marshall SJ (2021) Regime shifts in glacier and ice sheet response to climate change: examples from the Northern Hemisphere. *Front Clim* 3:702585. <https://doi.org/10.3389/fclim.2021.702585>
25. Martín-Moreno R, Allende-Álvarez F (2016) Little Ice Age glacier extension and retreat in Spitsbergen Island (High Arctic, Svalbard Archipelago). *Cuad Investig Geogr* 42(2):383–398. <https://doi.org/10.18172/cig.2919>
26. Miller GH, Geirsdóttir Á, Zhong Y et al (2012) Abrupt onset of the Little Ice Age triggered by volcanism and sustained by sea-ice/ocean feedbacks. *Geophys Res Lett* 39(2):L02708. <https://doi.org/10.1029/2011GL050168>
27. Ming J, Wang Y, Du Z, Zhang T, Wang W (2015) Widespread albedo decreasing and induced melting of Himalayan snow and ice in the early 21st century. *PLoS ONE* 10(6):e0126235. <https://doi.org/10.1371/journal.pone.0126235>
28. Owens MJ, Lockwood M, Hawkins E et al (2017) The Maunder Minimum and the Little Ice Age: an update from recent reconstructions and climate simulations. *J Space Weather Space Clim* 7:A33. <https://doi.org/10.1051/swsc/2017034>
29. PAGES 2k Consortium (2013) Continental-scale temperature variability during the past two millennia. *Nat Geosci* 6:339–346. <https://doi.org/10.1038/ngeo1797>
30. Pushpanjali B, Subrahmanyam MV, Murty KPRV (2013) Findlater Jet intensity and characteristics in relation to Indian summer monsoon. *Monsoons: formation, environmental monitoring and impact assessment*. Nova Science, New York, pp 47–63
31. Reinthaler J, Paul F (2025) Reconstructed glacier area and volume changes in the European Alps since the Little Ice Age. *Cryosphere* 19(2):753–767. <https://doi.org/10.5194/tc-19-753-2025>
32. RGI 7.0 Consortium (2023) Randolph Glacier Inventory – a dataset of global glacier outlines, version 7.0. NSIDC, Boulder. <https://doi.org/10.5067/f6jmovy5navz>
33. Roe GH, Baker MB (2014) Glacier response to climate perturbations: an accurate linear geometric model. *J Glaciol* 60(222):670–684
34. Salerno F, Guyennon N, Yang K et al (2023) Local cooling and drying induced by Himalayan glaciers under global warming. *Nat Geosci* 16(12):1120–1127. <https://doi.org/10.1038/s41561-023-01331-y>

35. Sampe T, Xie S-P (2010) Large-scale dynamics of the Meiyu-Baiu rainband: environmental forcing by the westerly jet. *J Clim* 23(1):113–134. <https://doi.org/10.1175/2009JCLI3128.1>
36. Sauter T, Brock BW, Collier E et al (2026) Glacier–atmosphere interactions and feedbacks in high-mountain regions – a review. *Rev Geophys* 64(1). <https://doi.org/10.1029/2024RG000869>. e2024RG000869
37. Schiemann R, Lüthi D, Schär C (2009) Seasonality and interannual variability of the westerly jet in the Tibetan Plateau region. *J Clim* 22(11):2940–2957. <https://doi.org/10.1175/2008JCLI2625.1>
38. Schurer AP, Tett SFB, Hegerl GC (2014) Small influence of solar variability on climate over the past millennium. *Nat Geosci* 7(2):104–108. <https://doi.org/10.1038/ngeo2040>
39. Schurer AP, Cowtan K, Hawkins E et al (2017) Importance of the pre-industrial baseline in determining the likelihood of exceeding the Paris limits. *Nat Clim Change* 7:563–567. <https://doi.org/10.1038/nclimate3345>
40. Steiner JF, Litt M, Stigter EE, Shea J, Bierkens MFP, Immerzeel WW (2018) The importance of turbulent fluxes in the surface energy balance of a debris-covered glacier in the Himalayas. *Front Earth Sci* 6:144. <https://doi.org/10.3389/feart.2018.00144>
41. Tao SY (1958) The association between Mei-yu in East Asia and seasonal variation of the general circulation of the atmosphere over Asia. *Acta Meteorol Sin* 29(2):119–134
42. Tielidze LG, Nosenko GA, Khromova TE et al (2025) Post-Little Ice Age equilibrium-line altitude and temperature changes in the Greater Caucasus based on small glaciers. *Remote Sens* 17(9):1486. <https://doi.org/10.3390/rs17091486>
43. Toropov PA, Debol'skii AV, Polyukhov AA, Shestakova AA, Popovnin VV, Drozdov ED (2023) Oerlemans minimal model as a possible instrument for describing mountain glaciation in Earth system models. *Water Resour* 50(5):675–687. <https://doi.org/10.1134/S0097807823700082>
44. Toropov PA, Shestakova AA, Muraviev AY, Drozdov ED, Poliukhov AA (2025) The new IGRICE model as a tool for studying the mechanisms of glacier retreat. *Climate* 13(12):248. <https://doi.org/10.3390/cli13120248>
45. Volodin EM, Mortikov EV, Kostykin SV et al (2017) Simulation of the present-day climate with the climate model INMCM5. *Clim Dyn* 49(11–12):3715–3734. <https://doi.org/10.1007/s00382-017-3539-7>
46. Volodin EM, Blagodatskikh DV, Bragina VV et al (2024) Computational framework for the Earth system modelling and the INM-CM6 climate model implemented on its base. *Russ J Numer Anal Math Model* 39(6):379–392. <https://doi.org/10.1515/rnam-2024-0032>
47. Wang B, Fan Z (1999) Choice of South Asian summer monsoon indices. *Bull Am Meteorol Soc* 80(4):629–638. [https://doi.org/10.1175/1520-0477\(1999\)080%3C0629:COASASM%3E2.0.CO;2](https://doi.org/10.1175/1520-0477(1999)080%3C0629:COASASM%3E2.0.CO;2)
48. Webster PJ, Yang S (1992) Monsoon and ENSO: selectively interactive systems. *Q J R Meteorol Soc* 118(507):877–926. <https://doi.org/10.1002/qj.49711850705>
49. Welch BL (1947) The Generalization of 'Student's' Problem when Several Different Population Variances are Involved. *Biometrika* 34(1–2):28–35. <https://doi.org/10.1093/biomet/34.1-2.28>

50. Wilks DS (2011) Statistical methods in the atmospheric sciences, 3rd edn. Academic, Oxford

51. Yeh TC (1958) The abrupt change of circulation over the Northern Hemisphere during June and October. *Acta Meteorol Sin* 29:249–263

52. Zasadni J (2007) The Little Ice Age in the Alps: its record in glacial deposits and rock glacier formation. *Stud Geomorphol Carpatho-Balc* 41:117–137

53. Zekollari H, Huss M, Farinotti D, Lhermitte S (2022) Ice-dynamical glacier evolution modeling—a review. *Rev Geophys* 60(2). <https://doi.org/10.1029/2021RG000754>. e2021RG000754

54. Zhang Y, Li T, Wang B (2004) Decadal change of the spring snow depth over the Tibetan Plateau: the associated circulation and influence on the East Asian summer monsoon. *J Clim* 17(14):2780–2793. [https://doi.org/10.1175/1520-0442\(2004\)017%3C2780:DCOTSS%3E2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017%3C2780:DCOTSS%3E2.0.CO;2)

Figures

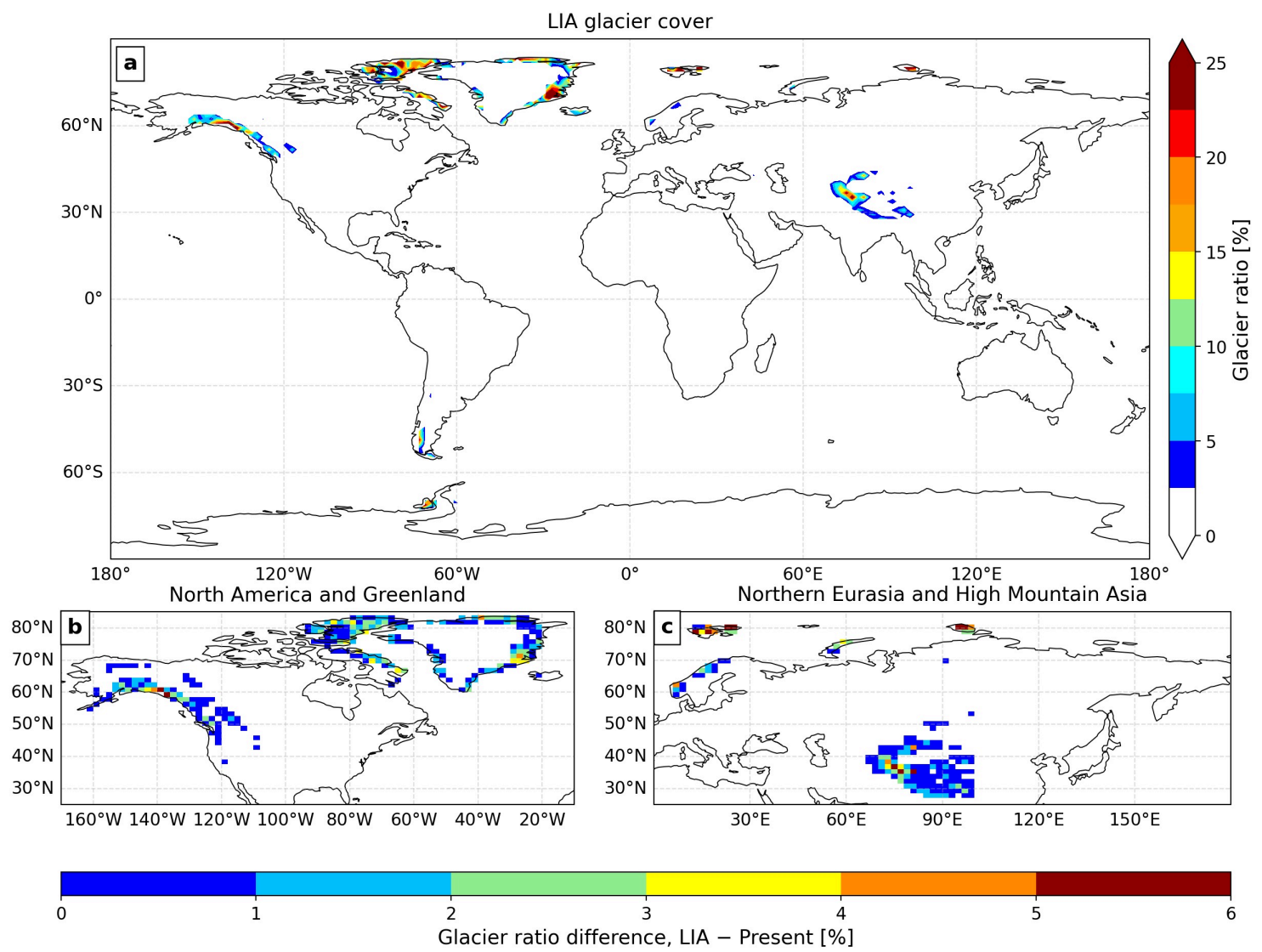


Figure 1

LIA mountain glacier mask prescribed in the glacier experiment. (a) Fractional glacier cover per grid cell at the LIA maximum extent (Sect. 2.2) - an absolute cover fraction, not a difference from the present day. (b, c) Difference between the LIA and present-day masks (LIA – Present), in percent of grid-cell area, over North America and Greenland (b) and Northern Eurasia and High Mountain Asia (c); both panels span 25-85°N. The difference is positive everywhere, hence the one-sided colour scale; note that the scales in (a) and (b, c) differ. Present-day glacier cover is from the Randolph Glacier Inventory (RGI Consortium 2017), regridded to the model grid.

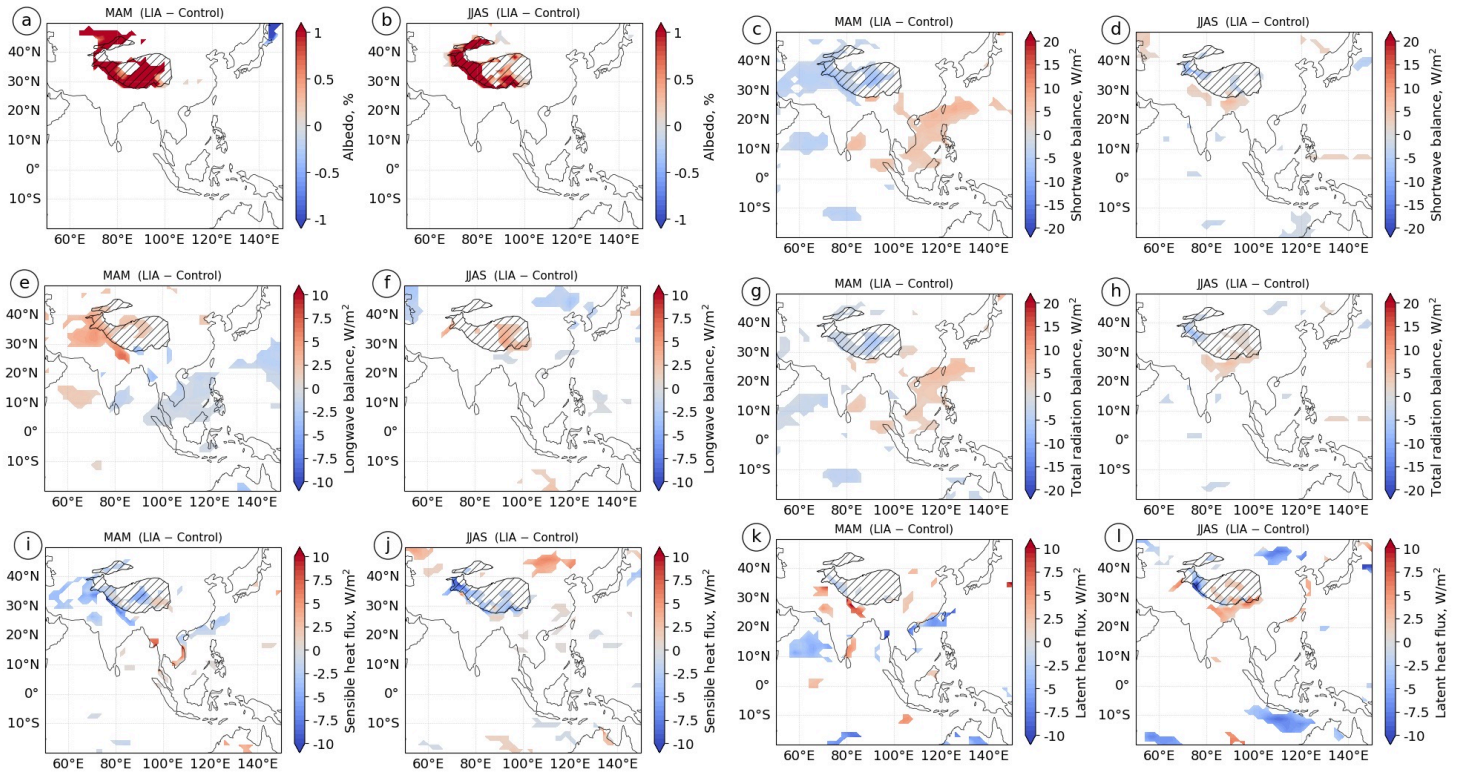


Figure 2

Anomalies (only significant ($p < 0.1$) are visible) for MAM and JJAS in Albedo, % (a, b); Shortwave balance, W/m² (c,d); Longwave balance, W/m² (e,f); Total radiation balance (g,h); Sensible heat flux, W/m² (i, j) ; Latent heat flux, W/m² (k, l). Tibetan Plateau is shown by the thick black contour on all panels.

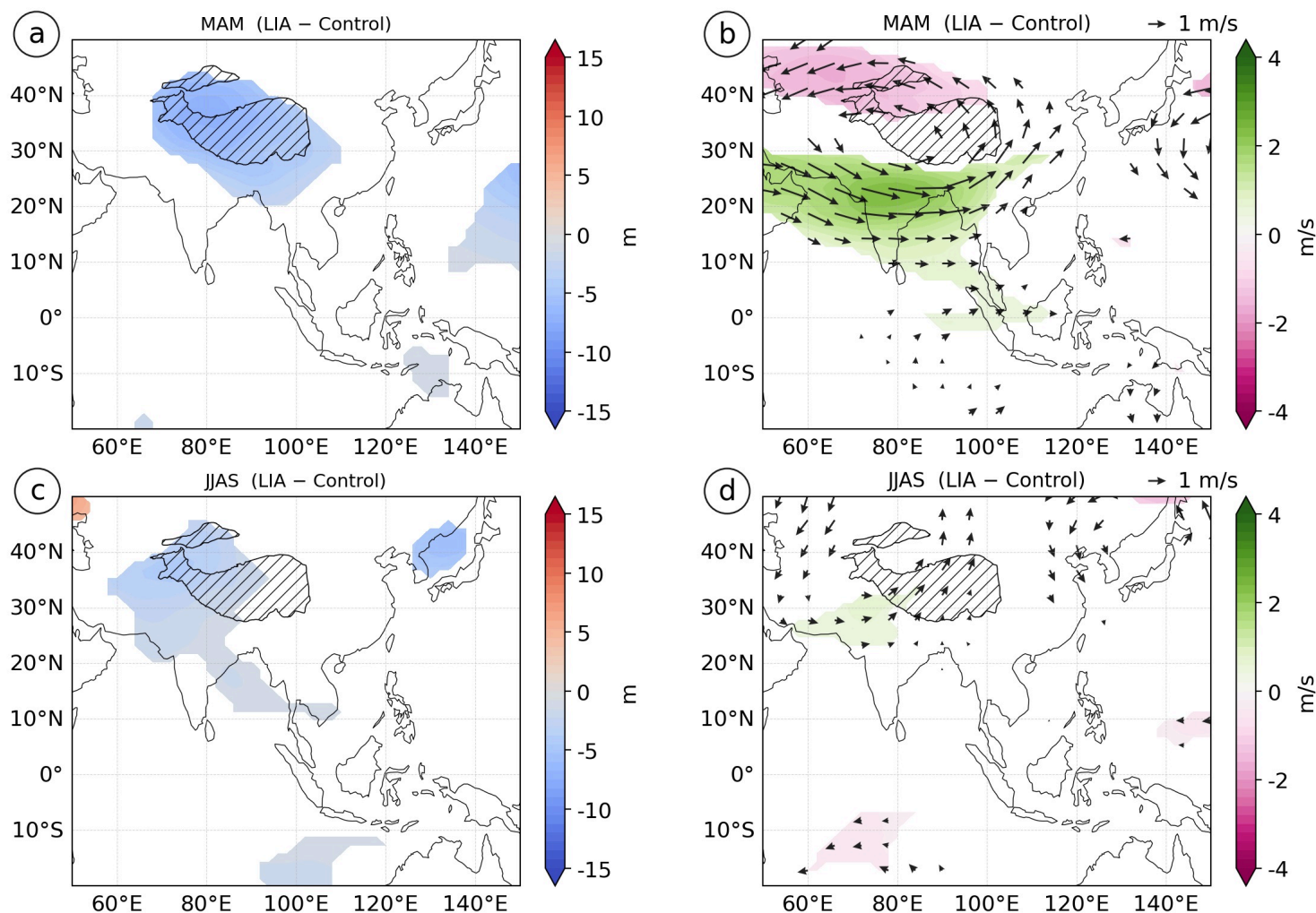


Figure 3

Geopotential height anomalies at 500 hPa in meters for MAM (a) and JJAS (c). Zonal wind anomalies at 200 hPa in m/s for MAM (b) and JJAS (d) overlaid with wind vector anomalies at 200 hPa as well (1 m/s reference arrow). Only statistically significant differences ($p < 0.1$) are shown.

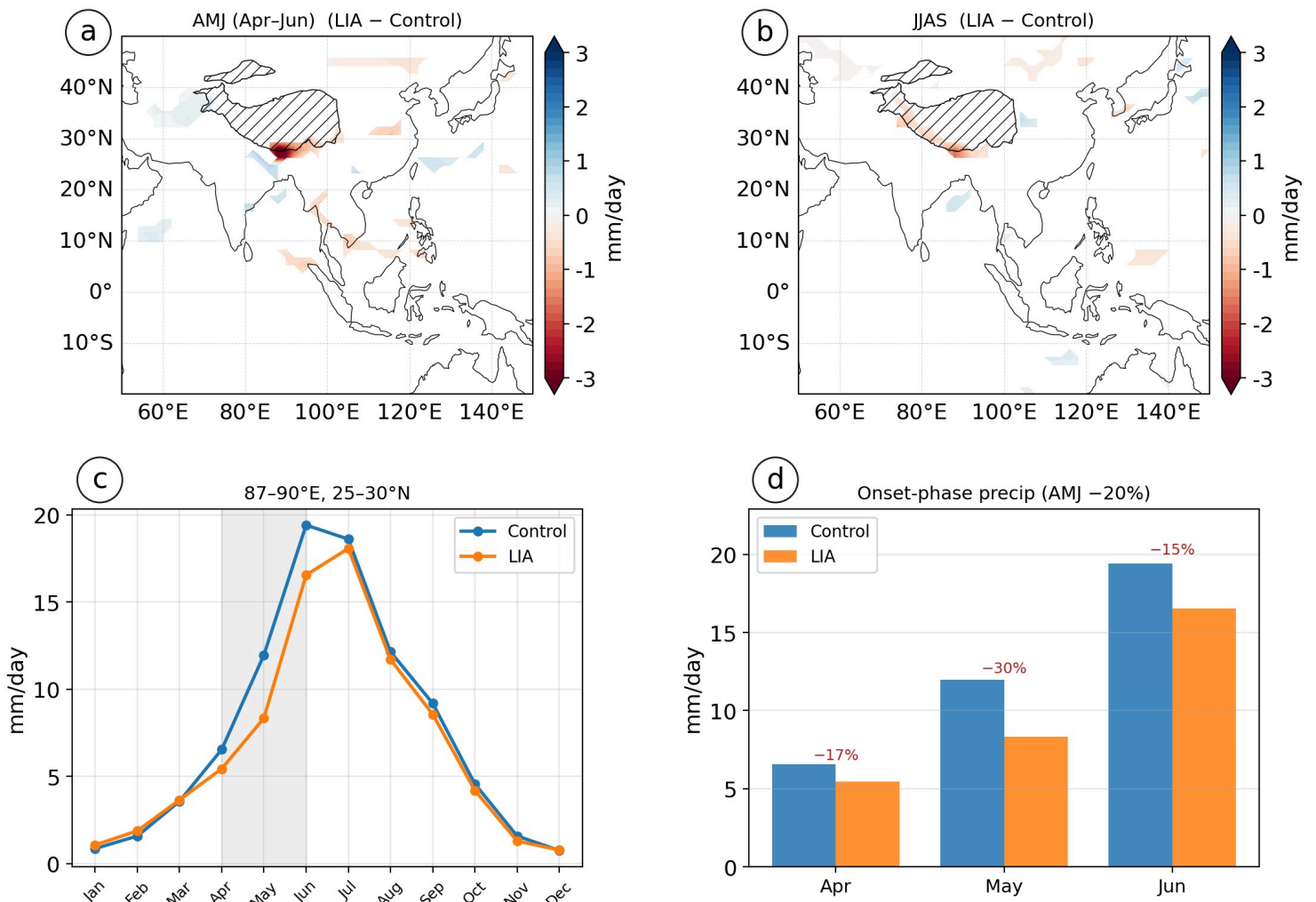


Figure 4

Model precipitation response (LIA - Control): (a) AMJ (April-June) anomaly (mm/day); (b) JJAS anomaly (mm/day); (c) annual cycle of area-mean precipitation over 87-90°E, 25-30°N (mm/day) for the Control and LIA experiments, with the April-June onset window shaded; (d) mean April-June precipitation by month for the two experiments, with the per-month percentage change annotated.

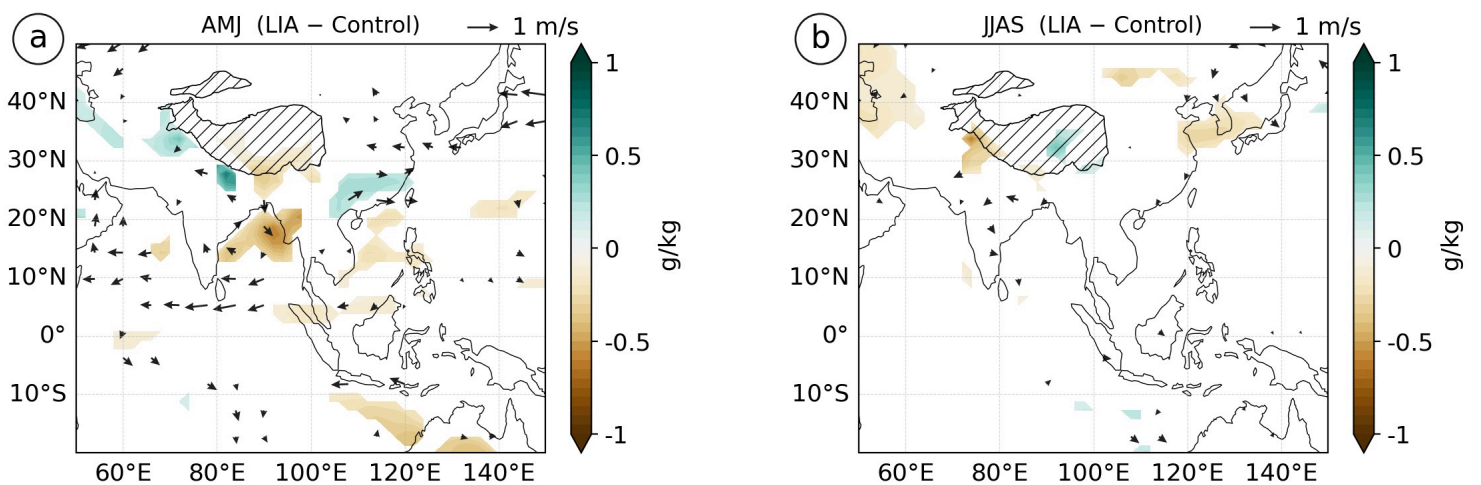


Figure 5

Atmospheric response to LIA glaciation (LIA minus Control). Specific humidity anomalies (g/kg, shading) at 850 hPa overlaid with wind vector anomalies at 850 hPa as well (1 m/s reference arrow) for (a) pre-monsoon (MAM) and (b) monsoon (JJAS) seasons. Only statistically significant ($p < 0.1$) anomalies are shown.