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# Spatiotemporal Heterogeneity in Cascading Snow-Hydrological Responses to Warming and Wetting on the Tibetan Plateau

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**Abstract.** Tibetan Plateau (TP) has experienced significant warming-wetting trends in recent years, inducing snowfall changes. However, the relative roles of precipitation and temperature in driving changes in snowfall and subsequent snow-hydrological processes remain poorly understood. We use China Meteorological Forcing Dataset (CMFD) v2.0 to quantify the contribution of meteorological factors including precipitation and air temperature to variations in snowfall. The results indicates that snowfall variability arises from complex interplay of both synergistic and competing effects between warming-wetting trends, with pronounced spatiotemporal and seasonal heterogeneity. Precipitation is the primary factor dominating snowfall variation during different stages, affecting snowfall changes across 66.42% areas of the TP. The compensatory effect of precipitation change offsetting warming-induced snowfall decline has intensified, especially after 2000, with positive contribution of 15.28 mm. Besides, the contribution mechanisms governing snowfall changes show phased elevation dependence, with mid-to-high elevation regions exhibiting greater sensitivity. Furthermore, sensitivity analyses based on the SSiB3 (Simplified Simple Biosphere Model version 3) land surface model reveal that warming-wetting induces cascading impacts through the "warming-wetting-snowfall-snow cover-snowmelt runoff (SMR)" chain. This cascading process leads to an overall increase of 2.2% in cumulative SMR, accompanied by enhanced spatial heterogeneity and seasonal variability of snowmelt. Distinct responses are observed among different snowfall-contribution regimes, snow regimes, and elevation bands. Changes in dominant climatic factors reshape the seasonal distribution of SMR. These findings reveal the complex mechanisms linking snowfall changes to cascading snow-hydrological responses under a warming-wetting climate over the TP, providing novel insights for water resource assessment and future climate adaptation.

## 1 Introduction

Snowfall is a sensitive indicator of climate change by modulating energy exchange between land and atmosphere through its high surface albedo and strong insulating capacity (Flanner et al., 2011; Jennings et al., 2018; Miao et al., 2024; Zhang, 2005). As the primary source of glaciers and seasonal snow, snowfall accumulates at the land surface to form snow storage and is gradually released as runoff through snowmelt processes. Therefore, changes in snowfall amount and its spatiotemporal distribution not only directly regulate snow accumulation but also reshape the timing, duration, and intensity of snowmelt

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(Zheng et al., 2022; Han et al., 2024). This further alters the generation mechanisms and seasonal distribution of snowmelt runoff (SMR). These cascading processes ultimately affect runoff regimes and hydrological stability, with important implications for water resources, ecosystems, and agricultural activities (Cho et al., 2021; Guo & Li, 2015; Li et al., 2025b).

Snowfall is jointly controlled by precipitation and temperature. Temperature governs precipitation phase by controlling the thermodynamic environment encountered by snowfall droplets during descent (Shi and Liu, 2024). Precipitation regimes primarily constrain the total amount of snowfall during snow season (Henderson et al., 2018; Mankin & Diffenbaugh, 2015). Observations indicate that global warming generally led to reduced snowfall by affecting precipitation phase transitions, particularly in the mid-to-low latitudes of the Northern Hemisphere (Lin & Chen, 2022; Shi & Liu, 2021; Siirila-Woodburn et al., 2021; Sun et al., 2024). However, increased seasonal snowfall due to greater atmospheric moisture capacity has been reported (Dai et al., 2008; Davis et al., 1999; Deng et al., 2017; Held & Soden 2006). The opposing impacts of temperature and precipitation on snowfall create significant challenges in interpreting changes in snowfall.

The Tibetan Plateau (TP), often considered as the “Third Pole”, features substantial permanent snow and ice cover in mid-latitudes (Immerzeel et al., 2010; Xu et al., 2019; Yao, 2019), exerting a significant impact on the water cycle in downstream regions and even globally (Chen et al., 2016; Zhu et al., 2017). As one of the most climate-sensitive regions, the TP has experienced pronounced “warming-wetting trends” over recent decades (Kuang & Jiao, 2016; Xu et al., 2020; You et al., 2020, 2021). The climatic changes have intensified precipitation phase transitions, exacerbating imbalance between solid and liquid water resources over the TP (Yao et al., 2022). Studies observed a general decreasing trend in snowfall over TP, specifically characterized by reduced snowfall in the southeast and increased snowfall in the northwest (Sun et al., 2024; Wang et al., 2016). However, other research highlights an increase in snowfall in the central and eastern regions of the plateau (Bai et al., 2019; Li et al., 2022; Zhang et al., 2024). Overall, estimates of plateau snowfall changes remain uncertain due to differences in datasets and climate models, while the complex topography further complicates conclusions about spatial variations in snowfall.

Numerous studies have employed correlation and trend analyses to elucidate the modulating effects of temperature and precipitation variations on snowfall changes over the TP. However, their conclusions exhibit significant regional heterogeneity and temporal variability (Wang et al., 2022; You et al., 2020). Climate model-based sensitivity experiments have also suggested that the reduction in snowfall fraction (Fs) caused by warming is the primary driver of snowfall loss. However, models struggle to capture the complex changes in meteorological variable, leading to large uncertainties in the results (Li et al., 2025a; Ma et al., 2023). Only a few studies have quantitatively assessed the contribution of precipitation variations to snowfall changes, but the complex effects of plateau warming and wetting on snowfall changes remains unclear (Dong & Ming, 2022; Gao et al., 2025). Moreover, previous studies mainly focus on changes in snowfall or snowmelt and their climatic attribution (Yang et al., 2022; Gao et al., 2023). However, how warming-wetting-induced snowfall changes propagate through snow accumulation and snow-hydrological process chains remains poorly quantified. This limits a comprehensive understanding of climate change signal transmission within coupled snow-hydrological systems.

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In this study, we characterize climatic changes over the TP using the China Meteorological Forcing Dataset (CMFD) v2.0. We further quantify the contributions of meteorological factors associated with snowfall changes, with a particular focus on precipitation- and temperature-driven components. The differential roles of precipitation and temperature are examined at both annual and monthly scales to better understand their temporal variability. In addition, we investigate the modulation of snowfall by elevation across the TP. Finally, we design a series of sensitivity experiments to explore snowfall changes under warming-wetting conditions and their impacts on snow accumulation and snow-hydrological processes.

## 2 Data and Methods

### 2.1 Meteorological forcing data

In this study, we use the China Meteorological Forcing Dataset v2.0 (CMFD v2.0, He et al., 2020) to assess the impact of warming-wetting trend on changes in snowfall over the TP. CMFD v2.0 provides high-resolution (0.1°) near-surface meteorological variables, including precipitation, temperature, surface pressure, etc., at 3-hourly intervals from 1951 to 2024. The dataset combines ERA5 reanalysis with ground-based observations and integrates AI-based radiation and precipitation products, substantially improving data accuracy relative to previous versions. Specifically, the dataset introduces precipitation products using Ding et al. (2014). to separate rain and snow beyond base variables. With extended temporal coverage, enhanced data quality and spatial consistency, CMFD v2.0 is well suited for climate, hydrology and ecological studies. All elements in CMFD v2.0 are selected from 1965 to 2020 for consistency, and interpolated on a spatial resolution of 0.1° (25°-40°N, 75°-105°E).

Furthermore, we employ a daily 0.05° snow depth (SD) dataset for TP (Yan et al., 2021) to delineate regions with thick snow (maximum SD ≥ 10 cm) and thin snow (maximum SD < 10 cm), while excluding areas with maximum SD below 2 cm. The dataset spans September 2000 to December 2021 at daily resolution. The product was generated by fusing a long-term Chinese snow depth dataset with a snow cover probability dataset using a sub-pixel spatiotemporal downscaling scheme applied to the original 0.25° snow depth fields. The downscaled product exhibits substantially improved accuracy, with the root mean square error (RMSE) reduced by 0.61 cm relative to the original dataset.

We also employ a 0.008° elevation dataset to calculate the grid terrain standard deviation. The dataset was constructed from National Aeronautics and Space Administration (NASA) Digital Elevation Model (DEM) data, incorporating terrain slope together with mountain integrity and ecosystem continuity. The DEM represents terrain relief as a gridded numerical surface and is widely used to characterize topographic variability, enabling detailed analyses of terrain controls on climate and cryosphere processes.

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### 2.2 Snowfall attribution analysis

In this section, we conduct an attribution analysis of recent snowfall change over the TP based on the methodology used in previous research (Gao et al., 2025; Krasting et al., 2013; Rasouli et al., 2022). Snowfall can be calculated as the product of total precipitation ($P$) and the snowfall fraction ($Fs$, the ratio of snowfall to total precipitation):

$$S = P \times Fs\ , \tag{1}$$

The change in snowfall ($\Delta S$) is defined as:

$$\Delta S = S_1 - S_0\ , \tag{2}$$

Where $S_1$ and $S_0$ are the average snowfall during the first five and last five years representing different research stages, respectively. Substituting Eq. (2) into Eq. (1):

$$\begin{aligned}\Delta S &= \Delta P Fs_0 + P\Delta Fs + \Delta P\Delta Fs \\ &= \Delta P\frac{Fs_0+Fs_1}{2} + \Delta Fs\frac{P_0+P_1}{2}\ ,\end{aligned} \tag{3}$$

Where $\Delta P\frac{Fs_0+Fs_1}{2}$ and $\Delta Fs\frac{P_0+P_1}{2}$ represent the contributions of total precipitation and snow fraction change, respectively. The variation of Fs is influenced by meteorological factors such as temperature, relative humidity, and surface air pressure (Ding et al., 2014; Jennings et al., 2018, Shi and Liu, 2024). Further, the relative contribution of different meteorological factors to changes in snowfall, through their influence on Fs, can be quantified using the elasticity method of Schaake (1990) and Zheng et al. (2009):

$$\Delta Fs = \varepsilon_T\frac{\Delta T}{\overline{T_{ref}}} + \varepsilon_{RH}\frac{\Delta RH}{\overline{RH_{ref}}} + \varepsilon_{P_s}\frac{\Delta P_s}{\overline{P_{s_{ref}}}}\ , \tag{4}$$

The coefficient $\varepsilon_{Var}$ can be calculated as $\frac{Fs_1-\overline{Fs_{ref}}}{\overline{Fs_{ref}}} = \varepsilon_{Var}\frac{Var_1-\overline{Var_{ref}}}{\overline{Var_{ref}}}$,where Var represents temperature ($T$), relative humidity ($RH$), and atmospheric pressure ($P_s$).

### 2.3 Seasonal distribution of snowmelt runoff

To quantify the seasonality of snowmelt runoff (SMR), two indices are used to characterize its timing and seasonal variability. Specifically, the center timing of SMR (CTSMR), representing the date when 50% of the annual SMR occurs (Court et al., 1962), is calculated as:

$$CTSMR = \frac{\sum_{i=1}^{365}(i\times q_i)}{\sum_{i=1}^{365}(q_i)}\ , \tag{5}$$

where i is time in days from the beginning of the hydrological year and $q_i$ is daily SMR (mm/day) for hydrological year day i. The SMR concentration index (SMRCI) is used to quantify the seasonal variability of SMR (Oliver et al., 1980) and is calculated as:

$$SMRCI = \frac{\sum_{i=1}^{365} {q_i}^2}{(\sum_{i=1}^{365} q_i)^2\times 100}\ , \tag{6}$$

The two SMR indices have been widely used in the literature to assess SMR seasonality characteristics under global change

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(Stewart et al., 2005).

### 2.4 Stage and typical contribution types classification

To minimize the impact of short-term fluctuations, this study employs a ten-year window method to analyze annual snowfall. This approach identifies three distinct turning points where differences in snowfall trends are most pronounced (Table S1 in Supporting Information S1), dividing plateau snowfall changes into three stages: Stage 1 (1966-1984), Stage 2 (1985-1999), and Stage 3 (2000-2020). Furthermore, based on the directional influence and relative contribution strength of precipitation and temperature on snowfall changes (Sect. 2.2), we categorize the factors affecting snowfall variations on the TP into four typical contribution types. Contributions where precipitation accounts for over 50% of snowfall changes are classified as precipitation-controlled (P-Ctrl), while those primarily driven by temperature are classified as temperature-controlled (T-Ctrl). Contributions where precipitation and temperature act in the same direction are classified as synergistic contributions, and those where they act in opposite directions are termed competitive contributions.

### 2.5 Model description and warming-wetting experiment design

The land surface model SSiB3-SNICAR (Oaida et al., 2015) is developed by coupling the snow albedo module SNICAR (Snow Ice and Aerosol Radiative; Flanner and Zender, 2005) into the SSiB3 model (Simplified Simple Biosphere Model version 3). The SSiB3 model includes the SAST snow scheme (Snow-Atmosphere-Soil Transfer model; Sun et al., 1999), which separates the thermodynamic processes of snow and soil and represents major snow processes, including snow compaction, heat conduction, snow grain growth, and snowmelt. The SNICAR module further improves snow albedo simulations by considering snow grain size evolution and aerosol effects on snow radiative processes. In SSiB3-SNICAR, snow is initially represented as a single thin layer. When snow depth exceeds 7 cm, the snow is divided into three layers. The model adopts the original SSiB3 snow cover fraction scheme (Sun and Xue, 2001) and applies the dynamic fresh snow density parameterization proposed by Anderson (1976), in which fresh snow density is controlled only by near-surface air temperature. To investigate the impacts of TP warming-wetting trends on snowfall changes and associated snow-hydrological processes, three sensitivity experiments are conducted using the SSiB3-SNICAR model (Table 1). The control experiment (CTL) uses the reference-period mean temperature ($T_{ref}$) and precipitation ($P_{ref}$) as inputs and applies the Ding method (Ding et al., 2014) to partition snowfall from precipitation. The reference period is selected as Stage 3, during which warming and wetting trends are most pronounced. Simulations are averaged over the first five hydrological years (2000.09.01-2005.08.31). In addition, the period from 1 January to 31 August 2000 is used as model spin-up.

To isolate the independent effects of warming and wetting, the warming experiment (EXP1) and the wetting experiment (EXP2) are designed. EXP1 imposes a temperature perturbation ($\Delta T$) relative to the reference temperature ($T_{ref}$), where $\Delta T$ represents the difference between the mean of the last five years in Stage 3 (2015.09.01-2020.08.31) and that of the reference period. In EXP2, the reference precipitation ($P_{ref}$) is multiplied by a scaling factor ($\delta P$), where $\delta P$ denotes the ratio of precipitation

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during the last five years of Stage 3 to that during the reference period. Both experiments use the Ding method for precipitation phase partitioning. Furthermore, the warming-wetting experiment (EXP3) simultaneously applies both temperature and precipitation perturbations to represent the coupled effects of warming and wetting. Specifically, under the wetting scenarios (EXP2 and EXP3), precipitation slightly decreases over parts of the southern TP but increases across most other regions, with an area-averaged increase of approximately 36.02 mm (Fig. 1). Under the warming scenarios (EXP1 and EXP3), air temperature increases across most of the TP, with the strongest warming occurring over the northwestern Plateau, where local warming reaches up to 2.5°C. In contrast, slight cooling appears along the southern margins, while the regional mean warming is approximately 0.58°C.

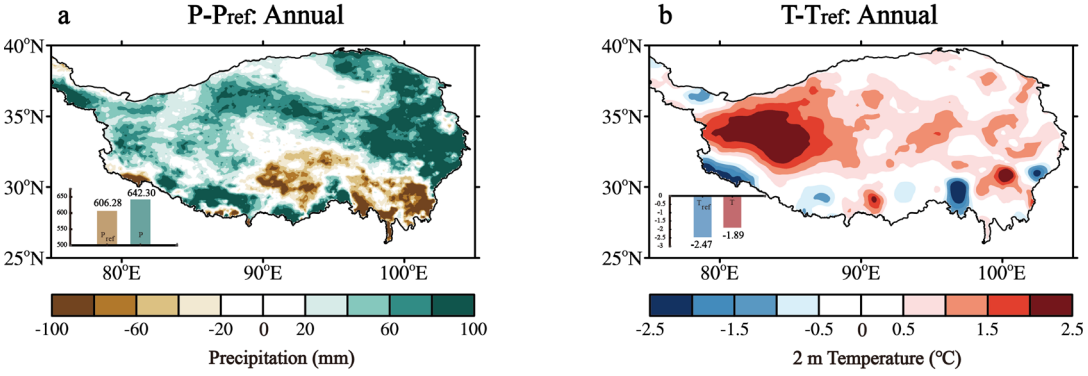



**Figure 1: Spatial distributions of warming and wetting perturbations during the Stage 3 (2000-2020) sensitivity experiments over the Tibetan Plateau.** Spatial distributions of changes in (a) precipitation ($\boldsymbol{P - P_{ref}}$) and (b) 2 m temperature ($\boldsymbol{T - T_{ref}}$) between the warming-wetting sensitivity experiments and the reference-period mean conditions. The lower-left corner in each panel shows the area-averaged value**.** Basemap source: Zhang et al. (2014).

**Table 1:** Experimental design of warming-wetting scenarios

| Experiments | | Temperature forcing (T) | Precipitation forcing (P) |
|---|---|---|---|
| CTL | Control experiment | $T_{ref}$ | $P_{ref}$ |
| EXP1 | Warming experiment | $T_{ref} + \Delta T$ | $P_{ref}$ |
| EXP2 | Wetting experiment | $T_{ref}$ | $P_{ref} \times \delta P$ |
| EXP3 | Warming-wetting experiment | $T_{ref} + \Delta T$ | $P_{ref} \times \delta P$ |

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## 3 Results

### 3.1 Warming-wetting trends and snowfall Changes over the TP

Climate change over the TP has experienced a pronounced phased transition. Figure 2m reveals a climatic shift in the mid-1980s, consistent with previous studies (Guo et al., 2012; Kang et al., 2010; Yao et al., 2022). Specifically, the precipitation trend reverses from a decreasing rate of -1.97 mm/year in Stage 1 to an increasing rate of 0.37 mm/year in Stage 2, while the warming rate intensifies from 0.02°C/year to 0.03°C/year (Table S1). The significant warming-wetting trend over the TP appears in the early 21st century. Precipitation shows a more significant increase at a rate of 1.66 mm/year. Concurrently, snowfall trends shift from sustained declines during Stage 1 (-2.6 mm/year) and Stage 2 (-1.01 mm/year) to a slight increase in Stage 3 (0.40 mm/year), while Fs does not increase synchronously. Given the pronounced warming and wetting during Stage 3 and the reversal in snowfall trends, this stage is selected as the focal point of the study to explore the impact of spatiotemporal heterogeneous warming and wetting on snowfall changes.

Spatially, precipitation trends have shifted from “east-wetting-west-drying” to “south-wetting-north-drying”, finally transforming into “south-drying-north-wetting” (Figs. 2a, e, i). Notably, the spatial patterns of snowfall and precipitation trends across different stages show a high degree of consistency, with both displaying a general expansion of their positive trend regions (Figs. 2g, k). Meanwhile, warming signals across parts of the plateau have intensified (Figs. 2b, f, j). Specifically, maximum precipitation increases of up to 5 mm/year occur over the eastern TP during Stage 3. Conversely, the western plateau presents a peak warming rate of 0.15°C/year. However, a significant cooling trend has emerged in the mountainous regions along the southern and western edges of TP. This cooling partially favors a transition from precipitation phase, contributing to an increase in Fs (Fig. 2l).

Overall, precipitation distribution primarily governs variations in snowfall amount, while the spatial variation in Fs trends is more closely linked to temperature. Notably, the Stage 3 indicates intensified warming and wetting over the TP. Meanwhile, precipitation trends exhibited a distinct north-south divergence, further accelerating the divergent snowfall changes. This highlights the combined effects of warming and wetting trends in shaping the complex response of snowfall over the TP.

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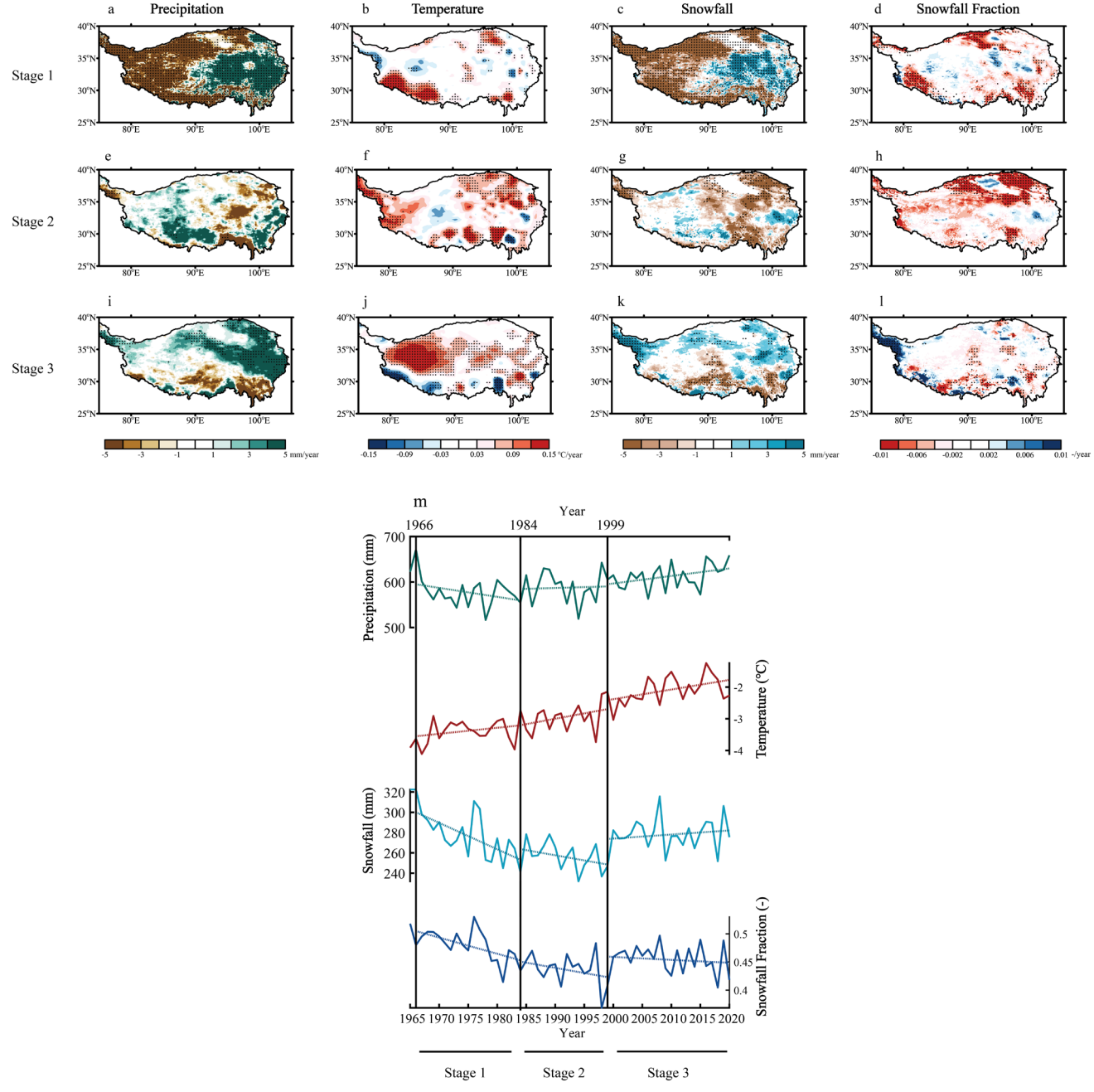



**Figure 2: Warming-wetting trends and snowfall change over the Tibetan Plateau during 1965-2020.** Annual spatial distributions of precipitation, temperature, snowfall, and snow fraction (Fs) during (a–d) Stage 1 (1966–1984), (e–h) Stage 2 (1985–1999), and (i–l) Stage 3 (2000–2020). (m) Corresponding interannual variations and linear trends. Dotted areas indicate regions where trends are statistically significant at the 95% confidence level. Basemap source: Zhang et al. (2014).

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### 3.2 Precipitation and temperature changes contribute to snowfall changes over the TP

To further analyze the impact of spatiotemporal warming and wetting trends on snowfall, we systematically quantify the contributions of different meteorological factors to snowfall variability (Sect. 2.2). Figures 3a, S1a, S2a reveal a significant shift in the spatial distribution of snowfall difference across three stages. Snowfall increases in the eastern TP and decreases in the west during Stage 1, whereas this pattern transforms to increased snowfall in the north and decreased snowfall in the south during Stage 3. This evolution closely follows the patterns of precipitation at the same time. The correlation coefficient between snowfall variations and precipitation changes reaches 0.47. In addition, the regional evolution of snowfall changes may affect the spatial distribution of snow accumulation and melt.

Further analysis reveals that the contribution from precipitation (C-Prcp) exhibits a distinct “strong-weak-strong” pattern across the three stages (Figs. 3b, S1b, S2b). The area of positive C-Prcp shifts progressively from the eastern TP to the central and northern regions as the plateau's wet zone transitions in phases, shaping the spatial distribution of increased snowfall. In contrast, the area characterized by negative contribution from snowfall fraction (C-Fs) expands continuously over time (Figs. 3c, S1c, S2c). This indicates a decline in the efficiency of precipitation phase transitions across much of the plateau, effectively suppressing snowfall formation. Notably, C-Prcp and C-Fs exhibit a pronounced competitive relationship during Stage 3 (Fig. 3d). The maximum positive C-Prcp exceeds 60 mm over the eastern TP. Nevertheless, C-Fs remains locally positive along the western margins and in parts of the southeastern TP, owing to regional temperature anomalies. Overall, snowfall on the TP shifts from significant decreases in Stage 1 (-37.65 mm) and Stage 2 (-16.36 mm) to a slight increase in Stage 3 (2.71 mm). Correspondingly, the influence of C-Prcp shifted from a negative effect (-24.15 mm) in Stage 1 to a positive effect (15.28 mm) in Stage 3. These results indicate that under intensified warming and wetting, precipitation changes increasingly favor snowfall enhancement over TP.

Further disassembly of C-Fs indicates that the increasingly negative C-Fs is primarily driven by rising temperatures. The competition is reflected in the growing magnitude of the temperature-driven negative component, C-Temp, which intensifies from -10.73 mm in Stage 1 to -18.84 mm in Stage 3 (Fig. 3h). Correspondingly, the areas with negative contributions due to warming are also expanding significantly (Figs. 3d, S1d, S2d). The strongest negative contributions are concentrated in the southeastern TP in Stage 3, exceeding 40 mm. In contrast, the spatial distribution of contribution from relative humidity (C-RH) and surface air pressure (C-Ps) is more dispersed and insignificant compared to C-Temp (Figs. 3e-f).

Overall, the pattern of snowfall difference changes in phases, closely following C-Prcp variations during different stages. Precipitation changes are the primary driver of snowfall variations. Conversely, warming exerts a negative regulatory effect, mainly through its influence on the phase transition of precipitation. In Stage 3, the compensatory effect of precipitation offsetting temperature has intensified, shifting from a synergistic to a competitive relationship. Furthermore, the plateau shows a high sensitivity to both precipitation and temperature changes, especially in the southeastern region. The significant spatial heterogeneity and competing effects between precipitation and temperature, contribute to the highly complex spatiotemporal

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evolution of snowfall changes across the TP. Furthermore, snowfall changes may influence snow accumulation and melt processes.

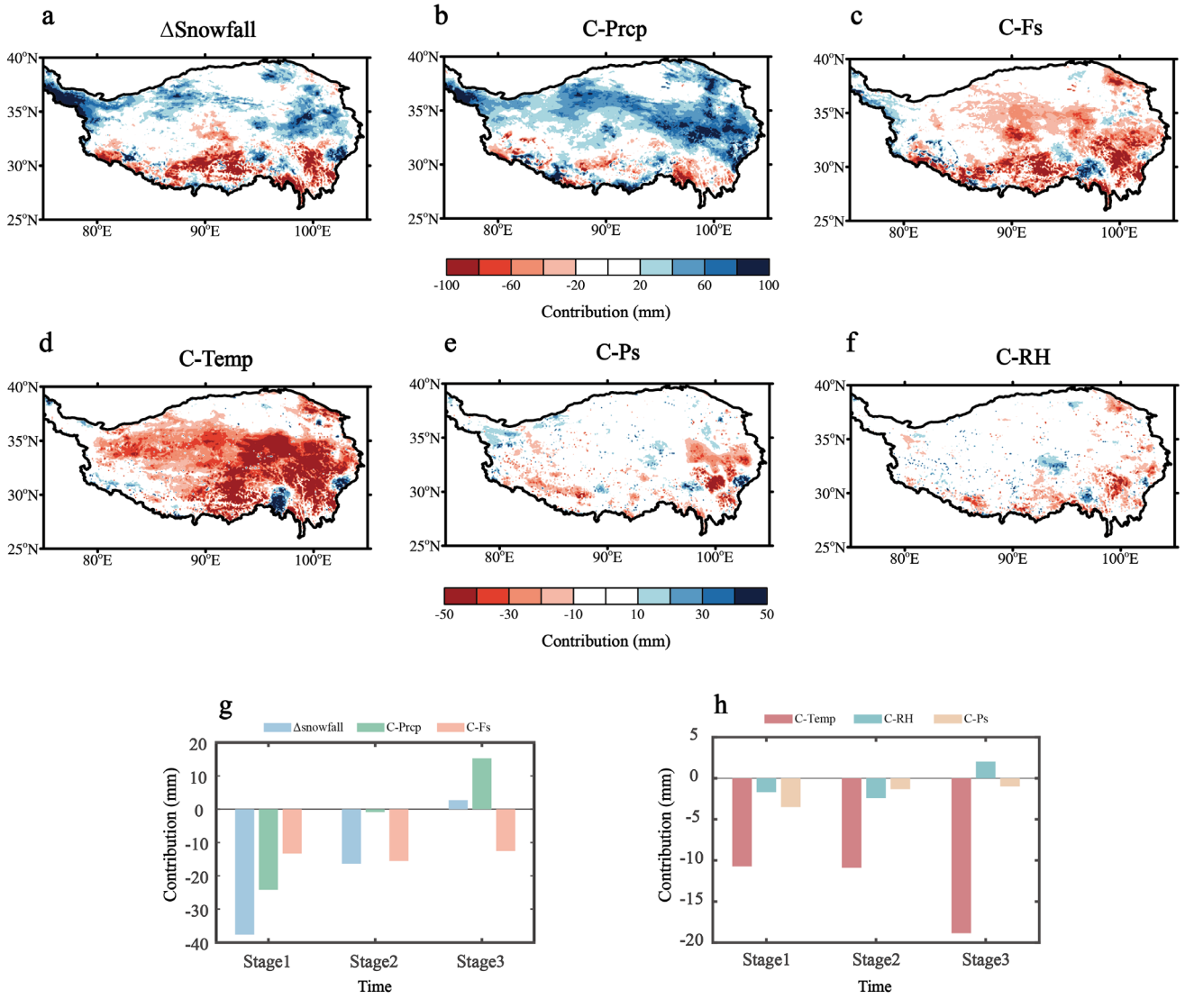



**Figure 3: Attribution of snowfall changes under the warming-wetting trend over the Tibetan Plateau (TP) during Stage 3 (2000-2020).** (a-c) Spatial patterns of the contribution from precipitation (C-Prcp), the snowfall fraction (C-Fs) to changes in snowfall (ΔSnowfall). (g) Comparison of the C-Prcp and C-Fs across the three stages. C-Fs is further decomposed into contribution from (d-f) temperature (C-Temp), relative humidity (C-RH), and surface pressure (C-Ps), with (h) comparing the contributions of three physical components across different stages. Basemap source: Zhang et al. (2014).

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### 3.3 Competitive effect between temperature and precipitation

To further investigate the contribution mechanisms of precipitation and temperature competition effects on snowfall variations on the TP, this study analyzes the dominant contributing factors influencing plateau snowfall variations at both annual and monthly scales. Figures 4m, S3m, and S4m indicate that the suppressive effect of rising temperature on snowfall strengthens progressively. However, P-Ctrl contribution remains the primary contributing type on the plateau. The zones of competitive contribution on the plateau have also experienced phased spatial migration and expansion. This migration corresponds to changes in the overlap between the wetting and warming zones (Fig. 2). Specifically, competitive regions increase markedly from 31.67% to 61.36%. The north-south contribution becomes more dominant in Stage 3. Annual snowfall variations in the northern TP are primarily P-Ctrl. By contrast, temperature exerts a stronger influence in the south. However, the proportion of P-Ctrl areas (66.42%) far exceeds that of T-Ctrl areas (33.58%).

Monthly analysis further reveals strong seasonal dependence and spatial heterogeneity in the relative roles of temperature and precipitation across different stages (Figs. 4a-l, S3a-l, S4a-l). During different stages, the temperature sensitivity of summer snowfall increases, while the P-Ctrl areas still dominate most regions. Additionally, competitive areas gradually expand and become more concentrated. P-Ctrl competitive regions prevail from late winter to early spring (January-March). In contrast, T-Ctrl competitive regions are concentrated in summer (July-August). However, a clear seasonal transition emerges in Stage 3. Summer snowfall variability shifts from P-Ctrl to T-Ctrl, with T-Ctrl areas reaching 52.59% in July and 68.64% in August. Spatially, the P-Ctrl and T-Ctrl competition contribution area in August is farther north than in July. Notably, the southern TP is characterized largely by synergistic effects. Meanwhile, the spatial distribution and seasonal evolution of contribution areas may further modulate the seasonal characteristics of snow accumulation and melt. Additionally, these patterns are closely linked to the non-uniform warming of the TP and the decadal-scale changes in large-scale atmospheric circulation. Some studies indicate that warming rates are higher in the north than in the south, particularly in summer (Duan et al., 2015). The strengthening of mid-latitude westerlies and the East Asian summer monsoon has also brought increased moisture to the region (de Kok et al., 2020). Meanwhile, the southern plateau is primarily affected by a decadal weakening South Asian summer monsoon (An et al., 2015; Liu et al., 2012; You et al., 2020; Zhang et al., 2019), where reduced precipitation combined with rising temperatures have led to a significant decrease in snowfall.

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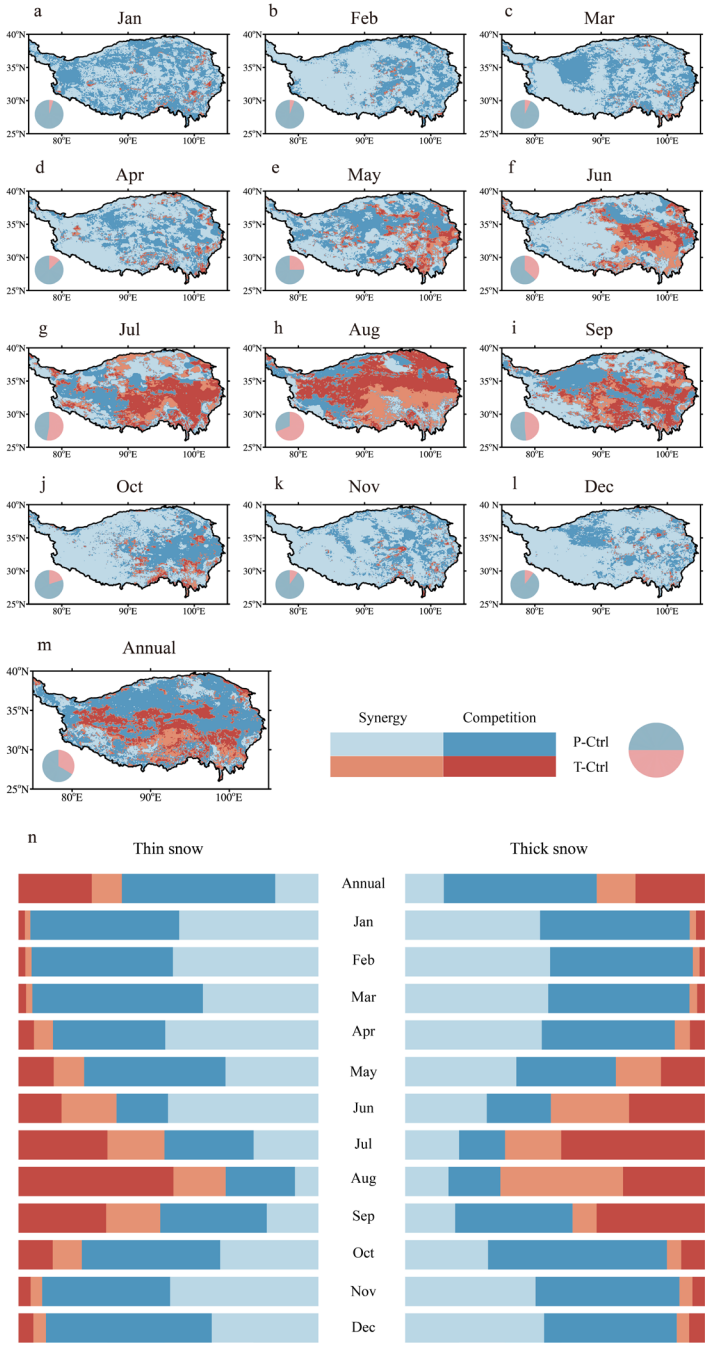



**Figure 4: Dominant contribution shifting in snowfall variations over the Tibetan Plateau (TP) during Stage 3 (2000-2020).** Distribution of controlling factors (precipitation and temperature) influencing (a-l) monthly and (m) annual snowfall variations, (n) with comparisons between thick and thin snow. The type of precipitation or temperature control (P-Ctrl or T-Ctrl) refers to precipitation or temperature contributing over 50% to snowfall variations. Synergistic and competitive types indicate identical or opposite contribution directions for both factors. Basemap source: Zhang et al. (2014).

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Owing to pronounced differences in snow depth, persistence, and associated local feedbacks, snowfall in different regions may exhibit distinct responses to changes in temperature and precipitation. At the annual scale, snowfall variability in both thin and thick snow regions remain primarily P-Ctrl across all stages, although the suppressive influence of warming intensifies progressively (Figs. 4n, S3n, S4n). With intensified warming and wetting, the fraction of T-Ctrl competitive regimes increases from 4.17% (2.45%) in thin snow (thick snow) regions during Stage 1 to 24.45% (9.79%) in Stage 3. Notably, P-Ctrl competitive regimes expand from 23.41% (17.31%) to 51.11% (21.55%). Thin snow regions show substantially greater wetting and warming sensitivity. Notably, the annual snowfall regime shifts from a predominantly synergistic to a competitive controlled structure in both snow regions during Stage 3, indicating an increasingly complex snowfall-forming environment.

Besides, summer snowfall shows markedly enhanced temperature sensitivity in thin-snow regions. During July-September, the cumulative fraction of T-Ctrl competitive regimes in thin-snow regions increases from 26.47% in Stage 1 to 110.46% in Stage 3. In thick-snow regions, the increase is more moderate, rising from 15.49% to 47.50%. By Stage 3, summer snowfall variability transitions from P-Ctrl to T-Ctrl. However, enhanced precipitation and associated moisture availability remain the dominant constraint on snowfall variability across both snow regimes.

Overall, under the warming-wetting trends of the TP, snowfall variations are influenced by both the synergistic and competitive effects of precipitation and temperature. The relative importance varies markedly across regions, seasons, and stages, which may further affect the seasonal distribution of snow-hydrological processes. Notably, P-Ctrl contributions dominate the long-term snowfall changes, while summer snowfall shows greater sensitivity to temperature. Moreover, the competitive influence between precipitation and temperature gradually intensifies and expands. Additionally, the regions of thin snow exhibit substantially greater sensitivity to warming and wetting trends than thick snow.

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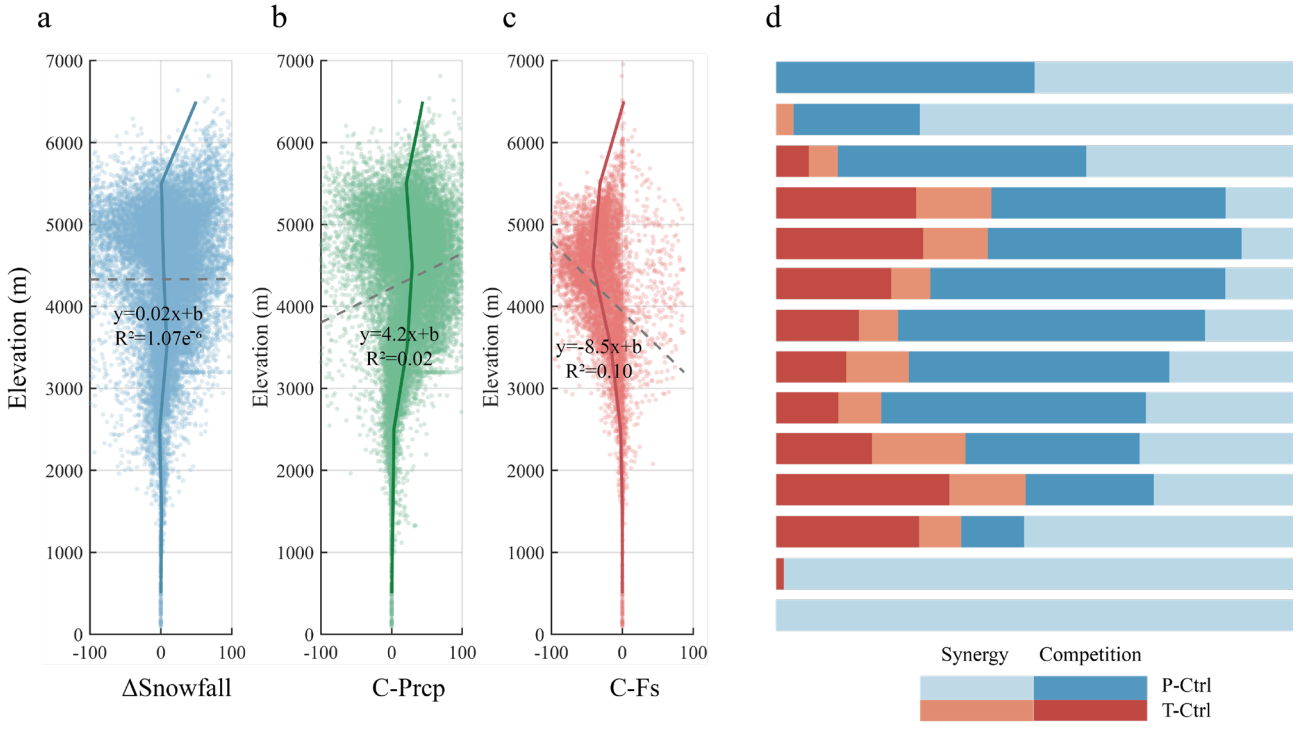



**Figure 5: Elevation-dependent modulation of precipitation and temperature contributions to snowfall changes over the Tibetan Plateau (TP) during Stage 3 (2000-2020).** (a-c) Variations in the contribution from precipitation (C-Prcp) and snowfall fraction (C-Fs) to changes in snowfall (ΔSnowfall) across increasing elevation, and (d) the dominant contribution types at different elevations are compared.

Influenced by complex terrain and multi-scale circulation patterns, snowfall variations over the TP exhibit distinct spatial heterogeneity and phased responses to warming trends. Figures S5a and S5e show that snowfall decreases with increasing elevation, displaying significant phased variations. The decrease in snowfall becomes more pronounced at elevations above 5000 m during Stage 1. Overall, negative C-Prcp dominates snowfall changes, with the P-Ctrl area significantly larger than T-Ctrl area (Fig. S5d). However, as the competing effects between warming and wetting intensify with time, the decline in snowfall across elevations gradually weakens and eventually reverses into increases at higher elevations (Figs. 5a-c).

In Stage 3, regions above 5000 m in western areas over the TP exhibits a clear rise in annual snowfall, while snowfall at mid-to-low elevations increases slightly (Fig. S8). This pattern reflects elevation-dependent variations driven by the interaction between C-Temp and C-Prcp. Between 4000 and 5000 m, sensitivity of snowfall to temperature increases, with C-Temp reaching -41.43 mm. This results in a competitive regime in which P-Ctrl areas exceed T-Ctrl (Fig. 5d). Specifically, T-Ctrl competitive zones account for 55.49% of the total area, while P-Ctrl competitive zones cover 94.55%. Gao et al. (2018) suggested that elevation-dependent warming (EDW) effect is more pronounced below 5000 m, influencing snowfall distribution. Enhanced water vapor availability at higher elevations explains the increased snowfall there. Mid-to-high elevations, corresponding to the TP snowline (Tang et al., 2014), show heightened temperature sensitivity.

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Overall, the contributions of precipitation and temperature to snowfall variations exhibit distinct elevation dependence and evolve across different stages. Specifically, the intensification of EDW and changes in the circulation background have complicated the competition between warming and wetting at mid-to-high elevations. This has amplified the heterogeneity of plateau snowfall changes. Notably, a reversal distinct from the first two stages emerged in Stage 3. Nevertheless, precipitation remains the dominant factor driving overall snowfall changes across all stages and elevation levels.

### 3.4 Snowmelt runoff responses under warming-wetting trends over the TP

To investigate the independent and coupled effects of warming and wetting on snowfall changes over the TP, a series of warming-wetting sensitivity experiments are conducted (Sect. 2.5). Overall, the differences in simulated snowfall among the experiments over the southern and northwestern TP are not fully propagated to the spatial patterns of the snow state variables, including snow water equivalent (SWE), snow depth (SD), and snow cover fraction (SCF) (Figs. 6a-i, S6a-f). Snow accumulation is strongly constrained by temperature. The lower temperatures over the southern TP likely favor a more efficient conversion of snowfall into snow. In contrast, variations in snowfall show a clearer correspondence with changes in SMR.

Relative to CTL, EXP1 (warming experiment) simulates an overall reduction in annual snowfall across most of the TP, except for the northwestern region, with a maximum decrease of approximately 10 mm/year. In contrast, the associated SWE anomalies exhibit opposite spatial patterns between the southern and northwestern TP. EXP1 further simulates an overall 4.5% reduction in cumulative SMR. In contrast, EXP2 (wetting experiment) exhibits an opposite snowfall response, with an overall increase of 2.3% in snowfall. This enhanced snowfall signal is further amplified in SWE and SMR, leading to an overall 4.7% increase in cumulative SMR. The snowfall change pattern simulated by EXP3 (warming-wetting experiment) is broadly consistent with the observed snowfall changes (Fig. 3a), with an overall decrease of 1.4%. Specifically, snowfall decreases over the southeastern TP and increases over the northern TP, with a spatial correlation coefficient of 0.29. This result suggests that the model reasonably reproduces snowfall changes under the warming-wetting climate background.

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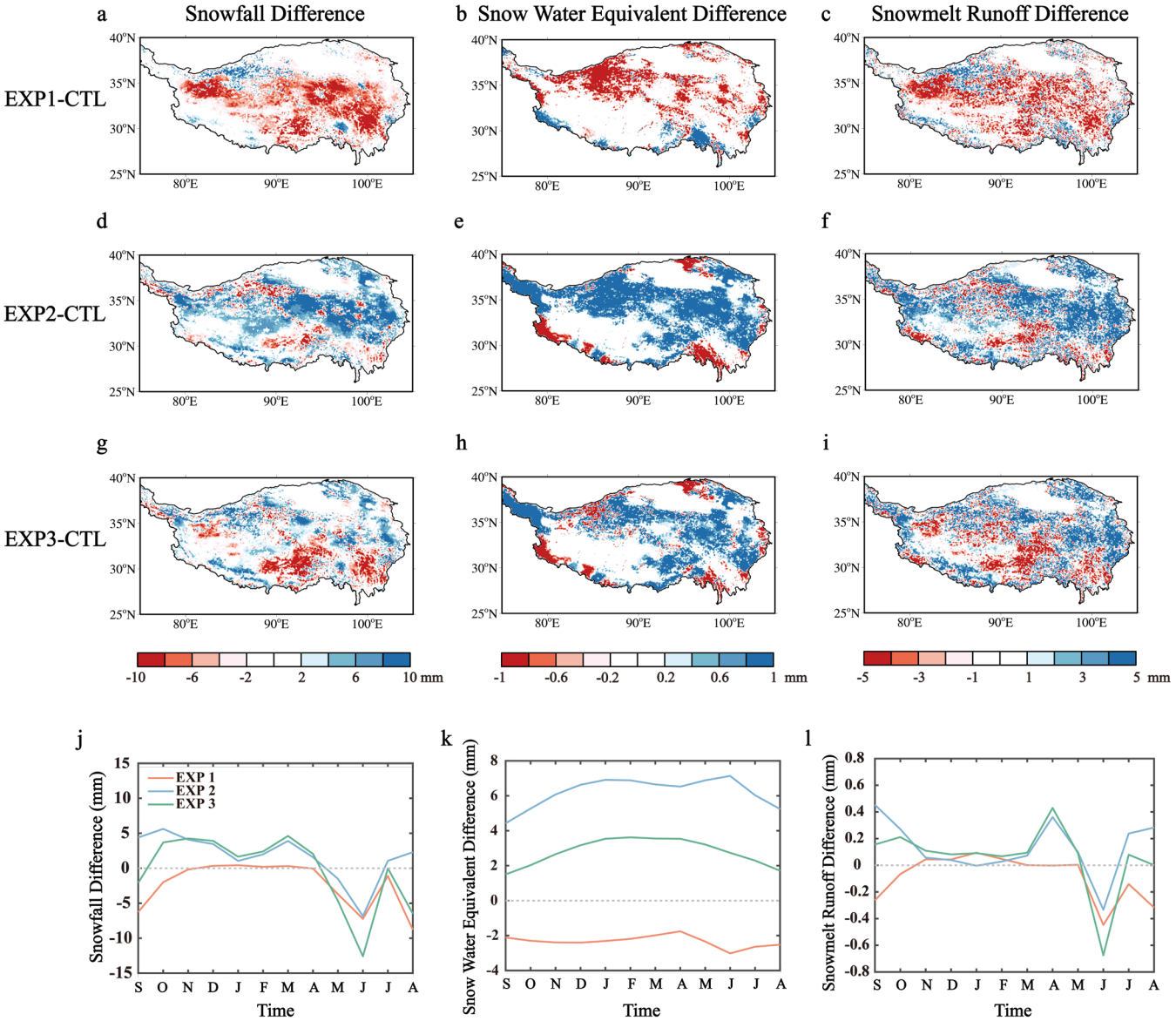



**Figure 6: Changes in snowfall and snow-hydrological processes in the three different warming-wetting scenarios for the Tibetan Plateau.** (a-c) Warming experiment (EXP1), (d-f) wetting experiment (EXP2), (g-i) Warming-wetting experiment (EXP3) and control experiment (CTL): simulated annual average (a, d, g) snowfall, (b, e, h) snow water equivalent and (c, f, i) cumulative snowmelt runoff, as well as (j-l) corresponding intra-annual circulation trends. Basemap source: Zhang et al. (2014).

Notably, EXP3 exhibits a snowfall difference pattern distinct from in EXP1 and EXP2, indicating that wetting effectively offsets the warming-induced reduction in snowfall. This spatial pattern is closely associated with the annual snowfall-contribution regimes classification, suggesting that snowfall changes over the TP are regulated by the synergistic and competitive effects of warming and wetting rather than by a single dominant factor. Furthermore, this pattern strongly influences the distribution of cumulative SMR, with amplified spatial contrasts and an overall increase of 2.2%. Meanwhile,

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the regional heterogeneity of snowfall changes further affects the seasonal distribution of cumulative SMR, as reflected by an earlier CTSMR in regions with reduced snowfall over the TP (Fig. S7e).

At seasonal scale, all experiments simulate substantially reduced snowfall in summer relative to CTL, with the largest decrease occurring in June and the strongest reduction found in EXP3 (Fig. 6j). In contrast, EXP2 and EXP3 both exhibits markedly increased snowfall during early spring and late autumn, whereas the differences between EXP1 and CTL remain relatively small during these periods. Notably, the summer and early autumn snowfall anomalies are further propagated to concurrent changes in cumulative SMR. Meanwhile, the pronounced increase in late-autumn snowfall enhances snow accumulation during the snow season and subsequently melts in the following spring. Together with increased spring snowfall, this process leads to a substantial enhancement of spring SMR, indicating a clear lagged "snowfall-snow cover-SMR" propagation effect (Figs. 6k-l). Specifically, the cumulative SMR simulated by EXP2 and EXP3 both peaks in April. Overall, the uneven seasonal redistribution of snowfall under different warming-wetting scenarios further intensifies the seasonality of SMR through its impacts on snow accumulation and ablation processes. Relative to CTL, the SMRCI increases by 1.29 and 0.77 in EXP2 and EXP3, respectively, whereas it decreases in EXP1.

The impacts of warming-wetting conditions on cumulative SMR over the TP exhibit pronounced regional heterogeneity, as the competitive and synergistic effects of precipitation and temperature reorganize the seasonal distribution of runoff. Distinct snowfall-contribution regimes exhibit pronounced seasonal spatial migration, causing their affected regions to shift throughout the year and leading to dynamically varying snowmelt runoff responses across the TP (Figs. 4a-l). In the P-Ctrl competitive regions (Fig. 7a), the wetting-induced increase in spring and summer snowfall in EXP2 and EXP3, relative to CTL, effectively compensates for the warming-induced reduction in snowfall and further enhances cumulative SMR. However, autumn warming reduces snowfall formation and consequently leads to lower cumulative SMR in EXP3 than in EXP2. In addition, the T-Ctrl regions show a stronger response to climate perturbations than the P-Ctrl regions. In the T-Ctrl competitive regions (Fig. 7c), Spring and summer warming offsets the wetting-induced increase in snowfall. Furthermore, the competitive effects between warming and wetting lead to insignificant changes in cumulative SMR simulated in EXP3. In contrast, winter warming accelerates snow ablation and significantly increases runoff generation. Notably, the synergistic effects of warming and drying on snowfall are particularly pronounced in summer across different control regions, jointly intensifying the substantial reduction in cumulative SMR (Figs. 7b, d). Overall, the spatiotemporal migration of different contribution regimes further enhances the regional heterogeneity of SMR responses under the warming-wetting background. Meanwhile, the relative dominance of warming and wetting reshapes snow-hydrological processes and their seasonal distribution by regulating snowfall changes.

Distinct SMR responses are also observed among different snow regimes (Fig. 7e). Overall, the thick- and thin-snow regions exhibit broadly consistent patterns of change, with both EXP2 and EXP3 simulating substantially enhanced cumulative SMR during early spring and late autumn. Compared with the thick-snow regions, the thin-snow regions show larger response magnitudes and asynchronous response timing. Specifically, SMR increases in EXP2 and EXP3 mainly occur in May and

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August within the thick-snow regions, whereas they occur in April and October within the thin-snow regions. This pattern indicates an earlier spring response but a delayed autumn response in the thin-snow regions.

Affected by the elevation-dependent warming-wetting characteristics over the TP, cumulative SMR also exhibits pronounced differences across elevation bands (Fig. 7f). Relative to CTL, EXP2 and EXP3 show similar vertical response structures. Specifically, within the mid-to-high elevation range (4000-5000 m), the increase in total SMR simulated by EXP2 is substantially larger than that in EXP3, reaching nearly twice the magnitude. Notably, EXP1 exhibits a clear response reversal within this elevation band, transitioning from increased SMR at lower elevations to reduced SMR at higher elevations. The magnitude of this change is comparable to that in EXP2, highlighting the suppressive effect of warming on high-elevation snowmelt processes.

Overall, warming-wetting conditions over the TP intensify the spatial heterogeneity of snowfall changes and further modify the spatiotemporal distribution of SMR by affecting snow-season accumulation. Notably, the cascade process of "warming-wetting-snowfall-snow cover-SMR" exhibits clear seasonal dependence. In late autumn, this propagation shows a lagged response due to the influence of winter snow accumulation, whereas in summer, snowfall variations directly control runoff changes. In addition, differences in dominant climatic controls across snowfall-contribution regions significantly regulate the seasonal redistribution of SMR. In P-Ctrl competitive regions, wetting strengthens the cascade during spring and summer, whereas in T-Ctrl regions, warming weakens it. Their combined effects intensify the reduction in SMR in summer. Furthermore, thin-snow regions exhibit an earlier spring response and a delayed autumn response of SMR compared with thick-snow regions. The elevation dependence of warming-wetting further enhances runoff heterogeneity, producing a runoff increase at mid-to-high elevations that differs from the response driven by warming alone.

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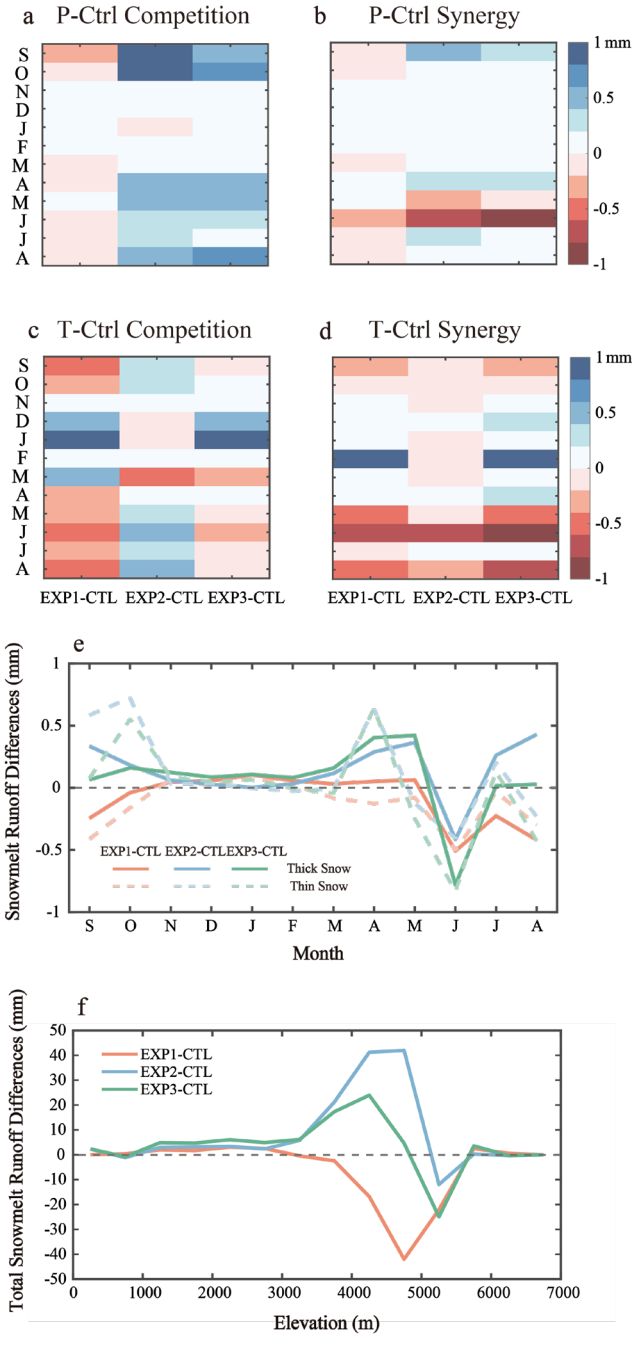



**Figure 7: Regional differences in simulated snowmelt runoff under different warming-wetting scenarios over the Tibetan Plateau.** Differences in monthly cumulative snowmelt runoff (SMR) between the warming experiment (EXP1), wetting experiment (EXP2), warming-wetting experiment (EXP3), and the control experiment (CTL) across different snowfall-contribution regimes, including (a-b) precipitation controlled (P-Ctrl) synergistic and competitive regions as well as (c-d) temperature controlled (T-Ctrl) regions. Differences in SMR are also compared (c) between thick- and thin-snow regions and (d) among different elevation bands based on total SMR.

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## 4 Conclusion and Discussions

This study uses CMFD v2.0 to quantify the snowfall responses to spatiotemporal variations under warming-wetting trend over the TP. Results show that temperature-related changes in snowfall fraction, Fs, primarily drive reduced snowfall, while precipitation remained the dominant factor influencing snowfall patterns. During 2000-2020, as warming and wetting intensifying, the effects of precipitation and temperature shift from synergistic to competitive regime, with precipitation contributing more positively by 15.28 mm.

Furthermore, the synergistic and competitive effects of precipitation and temperature on snowfall changes present significant spatiotemporal and seasonal differences. Precipitation is the main driver of snowfall changes, while summer snowfall is highly sensitive to warming. During 2000-2020, areas dominated by precipitation cover 66.42% of TP, while the area of competitive effects increases by 41.34% from 1966 to 2020. Notably, north-south differences emerge in the contributions of precipitation and temperature. Meanwhile, competitive effects amplify snowfall responses in thin-snow regions to the warming and wetting. Additionally, snowfall changes show an elevation dependence and evolve in phases, especially at mid-to-high elevations.

Finally, warming-wetting over the TP exerts a pronounced cascading influence on snow accumulation and snowmelt runoff (SMR) by altering the spatiotemporal distribution of snowfall. On the whole, cumulative SMR increases by 4.7% in wetting experiment and 2.2% in warming-wetting experiment relative to control experiment. Spatially, warming significantly weakens the wetting-induced increase in snowfall over the southeastern TP, while the northern TP is primarily dominated by wetting effects. Temporally, this influence exhibits clear seasonal differences, with a lagged response in autumn and a near-synchronous response in summer. Moreover, pronounced differences are found among snowfall-contribution regions, snow regimes, and elevation bands. Variations in dominant climatic controls strongly reshape the seasonal distribution of SMR. Together, warming and wetting over the TP exert cascading effects on snowmelt through the “warming-wetting-snowfall-snow cover-SMR” pathway, with pronounced regional and seasonal heterogeneity.

The evolution of TP warming-wetting trends warrants continued attention. Previous studies suggest that the plateau will continue to warm and generally become wetter in the future, albeit with strong spatial and seasonal heterogeneity (Hu et al., 2015; Zhao et al., 2022). These changes are closely linked to large-scale circulation adjustments. Under increasing greenhouse gas forcing, enhanced land-sea thermal contrast strengthens the South Asian summer monsoon, while intensified westerly moisture transport further increases atmospheric water vapor, contributing to enhanced humidity over the TP (Turner and Annamalai, 2012; Katzenberger et al., 2021). Meanwhile, greenhouse gas forcing, water vapor feedbacks, and snow-albedo interactions jointly accelerate warming over the TP, which exhibits pronounced seasonal and elevation-dependent characteristics (Huning & Margulis, 2018; Liu et al., 2020; Pepin et al., 2015; You et al., 2017, 2020).

Under these conditions, intensified warming-wetting is expected to reshape the regional hydrological cycle by modulating snowfall patterns and altering the role of seasonal snow as a water storage reservoir (Mernild et al., 2015; Hale et al., 2023). In this process, warming and wetting over the TP not only directly influence snowfall formation and snow evolution, but also generate interactive feedbacks through land surface processes such as snow variability, surface energy exchange, and soil

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freeze–thaw dynamics, making the snow-hydrological response no longer a simple one-way propagation. Consequently, snowmelt processes and their hydrological responses are governed by multi-process coupling, leading to enhanced interannual and seasonal variability in runoff (Berghuijs et al., 2014; Gordon et al., 2022; Ji & Yuan, 2024). Furthermore, snowfall changes may also influence glacier mass balance and permafrost degradation across the TP (Luo et al., 2025; Xiao et al., 2025). Therefore, hydrological changes under warming-wetting conditions are better characterized as a restructuring of hydrological processes and associated risk patterns rather than a simple increase in water availability. Overall, snowfall-driven cascading effects on snow-hydrological processes substantially enhance the variability of basin-scale hydrological systems, potentially increasing spring flood risk while exacerbating summer water shortages, thereby posing additional challenges to ecosystem stability and downstream socioeconomic systems (Chen et al., 2018; Han et al., 2024).

There remain uncertainties in this work. The elastic coefficient approach provides a first-order linear approximation of Fs intensity variations, without considering nonlinear interactions. The decomposition assumes factor independence, though in reality, large-scale climate change and local-scale meteorological interactions jointly influence snowfall patterns. However, this study focuses on primary snowfall drivers to ensure scientific validity. Moreover, the SSiB3-SNICAR model introduces certain uncertainties in representing snow processes, as simplifications in snow energy balance and parameterizations may affect snow accumulation and ablation simulations over complex terrain. Despite these limitations, the model remains suitable for capturing large-scale snow-hydrological responses over the TP. In addition, while CMFD v2.0 data contain uncertainties due to uneven station distribution and reanalysis errors, it remains the most reliable dataset for the TP. Future work can reduce uncertainties by incorporating more high-resolution, long-term observations, refining precipitation phase partitioning methods, and conducting higher-resolution land-atmosphere coupled simulations.

## Data availability

The China meteorological forcing dataset v2.0 (CMFD V2.0) is downloaded at https://data.tpdc.ac.cn/zh-hans/data/e60dfd96-5fd8-493f-beae-e8e5d24dece4. And the 0.05° snow depth dataset for TP is available at https://data.tpdc.ac.cn/en/data/0515ce19-5a69-4f86-822b-330aa11e2a28/. The 0.008° elevation dataset is available at https://www.tpdc.ac.cn/zh-hans/data/ddf4108a-d940-47ad-b25c-03666275c83a/.

## Author contributions

Xinyun Hu: Writing - original draft, Methodology, Formal analysis, Data curation.
Xin Miao: Writing - review & editing, Methodology, Supervision.
Shangyong Shi: Writing - review & editing.
Le Wang: Writing - review & editing.
Yizhuo Li: Writing - review & editing.

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Qing Guo: Writing - review & editing.

Weidong Guo: Writing - review & editing, Supervision.

## Competing interests

The authors declare that they have no conflict of interest.

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