Potential links between wintertime snow cover in central Europe and precipitation over the low-latitude highlands of China in May

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Abstract
This research describes a link between winter–springtime snow cover anomalies in central Europe and precipitation over the low-latitude highlands of China (LLHC). Excessive snow cover over Europe and East Asia alters the meridional temperature gradient, which induces zonal wind anomalies at the 500 hPa level that impact the strength and position of subtropical streams. These anomalies persist until spring and result in a distinctive paired cyclonic–anticyclonic circulation pattern. Over Mongolia, the imposition of cyclonic circulation weakens the Mongolia High and reduces the potential southward flow of cold air that normally affects the LLHC during spring. Concurrently, the development of anticyclonic conditions over the LLHC weakens springtime precipitation in this area. This paired circulation system is verified by the linear baroclinic model, which also confirms that the observed linkage between snow cover and precipitation is not modulated by the El Niño–Southern Oscillation. This analysis affords new insight into the effects of snow cover on the subtropical stream and atmospheric circulations downwind.

KEYWORDS
low-latitude highlands of China, precipitation, snow cover

1 | INTRODUCTION

Snow cover plays a significant role in the climate system. By exerting a cooling effect on local land surfaces (Flanner et al., 2011), snow can impact the vertical structure of the overlying atmosphere and influence conditions farther downstream via a chain of subsequent atmospheric circulations. At ground level, temperature changes related to snow cover can also be principal drivers of extratropical climate change (Serreze and Barry, 2011; Qu and Hall, 2014). Meanwhile, at lower latitudes, snowmelt serves to increase soil moisture levels and cool the ground surface, thereby reducing evaporation (Yeh et al., 1983). This long-term process is a key component of climate variability in tropical regions (Qian et al., 2003).

Since the foundational work of Cohen and Entekhabi (1999), numerous observational and modeling studies have sought to establish whether variability in Eurasian snow cover during the autumn can perturb Northern Hemisphere atmospheric circulation via the generation of deep, planetary atmospheric waves (Henderson et al., 2018). Such a mechanism, first proposed by Saito et al. (2001), involves strengthening of the
Siberian High, upward propagation of a Rossby wave into the polar stratosphere, and the subsequent downward propagation of stratospheric circulation into the troposphere. In a related scenario, anomalous snow cover over the Tibetan Plateau is implicated in fluctuations of the Indian summer monsoon; elevated winter–spring snowfall in Tibet is typically followed by weakening of the monsoon and drier conditions over India (Blanford, 1884). Subsequent studies have revealed that Tibetan Plateau snow cover is not only a strong indicator of Indian Summer Monsoon strength (Hahn and Shukla, 1976; Zhao and Moore, 2004; Yuan et al., 2019), it also impacts precipitation in East Asia and the East Asian Summer Monsoon (Qian et al., 2003; Liu et al., 2014; Xiao and Duan, 2016; You et al., 2020).

Yet, despite the fact that snow cover is a demonstrated climatic parameter, its impact on low-latitude highland precipitation has not been investigated fully. The low-latitude highlands of China (LLHC) are located between 30°S and 30°N and have a mean elevation exceeding 1,500 m (Tao et al., 2013). Positioned with the Bay of Bengal to the southwest, the South China Sea to the southeast, and the Tibetan Plateau to the northwest, the LLHC experiences a complicated precipitation regime (Figure S1; Cao et al., 2014). Previous studies have focused largely on modulation of LLHC precipitation by sea surface temperature (SST), the Indian Summer Monsoon, and the East Asian Summer Monsoon; the impact of snow cover has not been investigated (Cao et al., 2014, 2017; Tao et al., 2016; Wang et al., 2018). Therefore, this research explores the potential linkage between Eurasian snow cover and springtime (May) precipitation over the LLHC. Specifically, we address the research questions: (a) How does Eurasian snow cover relate to LLHC precipitation? (b) What are the dynamic processes linking the two and what are the implications for atmospheric circulations downstream, with particular emphasis on May precipitation over the LLHC?

2 | ANALYSIS OF RESULTS

2.1 | Potential linkage between Eurasian snow cover and May rainfall over the LLHC

For our singular value decomposition (SVD) analysis, we employed 3-month averaged Eurasian snow cover extent (SCE) data derived from NCEP–NCAR reanalysis (left field) for boreal winter (December–January–February; DJF) and spring (March–April–May; MAM), in conjunction with precipitation data from 209 stations located throughout the LLHC (right field). The SCE data were weighted by latitude and the Nino3 component removed prior to analysis. We also excluded the negative trend of spring snow cover due to global warming (Brown and Robinson, 2011) to isolate and evaluate the relationship between snow cover and precipitation.

We found that the leading SVD mode explains ~43.2% of the total variance at >95% confidence, indicating a close relationship between Eurasian snow cover and May precipitation over the LLHC (Figure 1). Specifically, extended snow coverage over central Europe correlates with decreased rainfall over the LLHC. There is a small seasonal variability in the range and position of key snow-covered regions, reflecting relative seasonal stability of the area where the homogeneous coefficient exceeds 99% confidence (Figure 1a–d). Correspondingly, precipitation over the LLHC responds with a robust negative signal. An extensive portion of the LLHC passed the 99% confidence level, while the remainder exhibits the same symbolic coefficient (Figure 1e).

Time series for the left (snow cover) and right (rainfall) fields return a correlation coefficient of 0.77 (>99% confidence) and account for >59% of the observed precipitation change over the LLHC. Our analysis confirms that LLHC rainfall is modulated by the Indian and East Asian Summer Monsoons (Cao et al., 2012, 2014), and indicates that the El Niño–Southern Oscillation (ENSO) plays a key role in bridging Eurasian winter snow cover and monsoon intensity (Yang, 1996). In contrast to snow-related connections reported by previous studies, our data also show that the linkage between Eurasian snow cover and LLHC precipitation is relatively robust and independent of ENSO modulation.

The independent nature of this linkage aligns with the findings of earlier studies. However, whereas previous researchers sought to explain variability in LLHC precipitation by SSTs, particularly the strong ENSO signal, our findings reveal that, despite the Pacific basin experiencing contrasting SST patterns during the period 2009–2011 (El Niño in 2009 and La Niña in 2011), the LLHC nonetheless experienced similarly negative rainfall anomalies. This outcome indicates that the external SST forcing (especially ENSO alone cannot explain the observed variations in May rainfall over the LLHC (Cao et al., 2014). Furthermore, previous work by Chang et al. (2001) reported a pronounced weakening of the relationship between ENSO and Indian Monsoon rainfall since the 1970s, potentially reflecting the alteration of meridional temperature gradients by shifting winter surface temperatures over Europe, similar to the linkage we propose here. Together, this prior work supports the hypothesis that May precipitation over the LLHC is influenced by factors other than ENSO and suggests that our proposed relationship might also operate
FIGURE 1  Results of the leading mode of SVD analysis with ENSO signal removed. (a–d) Homogeneous coefficient of the left field; (e) heterogeneous coefficient of the right field; (f) normalized expansion coefficient. Graduated red (blue) shading in (a–e) indicates areas of positive (negative) anomalies that are significant at the 90% (light shading), 95% (medium shading), and 99% (heavy shading) confidence levels. Black dots in (e) represent the locations of the 209 stations providing precipitation data. Red (blue) line in (f) represents the homogenous expansion coefficients of the left field (right field), and red (blue) dots indicate the homogenous expansion coefficients of the left field (right field) for each year. The variance contribution of the first SVD mode is given in the top-right of the figure. The correlation coefficient between the left and right field time series is given in Figure S2f

TABLE 1  Regionally averaged composite surface temperature anomalies (units: °C) and surface fluxes anomalies (units: W m⁻²) over the key region (10°–70°E, 45°–55°N) of seasonal Eurasian snow cover

|                  | DJF   | JFM   | FMA   | MAM   |
|------------------|-------|-------|-------|-------|
| Temperature      | −3.27 | −3.59 | −2.90 | −0.95 |
| Surface upward longwave radiation | −13.31 | −15.05 | −12.52 | −4.09 |
| Surface upward shortwave radiation | 3.42  | 5.72  | 6.89  | 4.92  |
| Surface latent heat flux | −1.21 | −3.47 | −4.23 | −3.27 |

Note: Surface fluxes include surface upward longwave radiation, surface upward shortwave radiation, and surface latent heat flux.
independent of ENSO modulation. To confirm the importance of snow cover, we performed a simple linear regression (Table S1) that shows snow cover is a stronger statistical predictor than ENSO (Figure S5).

To investigate the influence of land surface processes on snow cover, and therefore on LLHC precipitation, we conducted a composite analysis of regionally averaged surface temperature and relevant surface fluxes for the principal region of Eurasian snow cover (10°–70°E, 45°–55°N; Table 1), where the snow cover coefficient exceeds the 99% confidence level (Figure 1a–d). We picked the years with higher than 0.8 standard deviation of the snow

![Composite of the zonal wind anomalies averaged over 10°–70°E (units: m·s⁻¹) in (a, c, e, g) reanalysis data and (b, d, f, h) simulated zonal wind anomalies in the LBM model. Dotted areas are significant at the 95% level. Black contours demark the climatological zonal wind, CI = 10 m·s⁻¹](image-url)
cover time series provided by the leading SVD mode to conduct composite analysis. Therefore, we selected a total of 10 positive-anomaly years (1979, 1980, 1982, 1985, 1987, 2003, 2005, 2010, 2011, and 2012) and 11 negative-anomaly years (1981, 1989, 1990, 1993, 1995, 2000, 2001, 2002, 2004, 2007, and 2016).

Table 1 shows that extensive snow cover corresponds to abnormally negative surface temperatures and anomalously low fluxes of upward longwave radiation from the surface, resulting in a strong cooling effect on the overlying atmosphere. We observed that the temperature and surface flux anomalies are greatest in winter and late winter (January–February–March; JFM), and decline over time thereafter. When the surface latent heat flux was greatest in early spring (February–April; FMA) and spring, this was due to the extensive melting of snow at that time. However, we note that the extent of Eurasian snow cover is relatively small in spring, particularly in the key region for our study.

### 2.2 A possible mechanism for the snow–precipitation connection

As illustrated by Figure 2a–d, our composite analysis reveals that extensive snow cover alters both the strength and position of zonal winds, with prescribed wind anomalies prevalent along the climatological high-wind stream. Specifically, we found that the subtropical stream both strengthens and shifts northward of its mean position. One explanation for these anomalies is that they reflect strengthening of the meridional temperature gradient in subtropical latitudes due to the cooling effect of snow, since the core of the abnormal zonal winds coincides geographically with the key region of snow cover. This cooling effect reduces surface air temperatures (Table 1), which in turn strengthens the subtropical meridional temperature gradient, resulting in the observed zonal wind anomalies. This scenario aligns with the theory of thermal wind (Figure 2a–d) and might also serve to intensify and displace the subtropical high-wind stream. We note that these two zonal wind anomalies can persist until spring, when the area of snow cover and its associated cooling effect diminish.

The abnormal zonal winds not only characterize the key region of snow cover, but also propagate along the stream (Figure 2a–d), resulting in the acceleration and northward displacement of the subtropical high-wind stream. Westerly wind anomalies at 30°N and easterly anomalies near 60°N, together with the low-pressure zone separating them, combine to form a robust circulation that persists until spring (Figure 2). As shown in Figure 2, anomalous snow cover alters the meridional temperature gradient, resulting in the development of a cold low-pressure system over Europe and West Asia and high-pressure conditions farther north. In springtime, this circulation pattern results in westerly wind anomalies at 30°N, easterly wind anomalies at 60°N, cyclonic circulation over Mongolia, and anticyclonic circulation over the LLHC (Figure 2d).

We note that the cyclonic circulation over Mongolia is sustained until spring by the two zonal wind anomalies depicted in Figure 4. These anomalies together comprise a secondary circulation located to the north of the stream and potentially associated with the relatively weak, yet persistent, anticyclone over the LLHC. We suggest that this subsequent circulation weakens the Mongolia High, thereby inhibiting the southward passage of cold air from Siberia; the concurrent anticyclonic circulation over the LLHC causes divergence of water vapor from the Bay of Bengal. Ultimately, this secondary circulation inhibits the supply of precipitation to the LLHC.

These secondary atmospheric conditions can also be evaluated from the horizontal distribution of vertical air velocity (Figure S4). Employing the 500 hPa level (non-divergence level), we investigated the distribution of the vertical-velocity anomaly composite and found that an abnormal downdraft exists over the LLHC, contributing to atmospheric stability and the negative trend of regional rainfall.

In summary, our findings suggest that cold temperature anomalies (Table 1) extending from excessive snow cover strengthen the subtropical meridional temperature gradient, intensify the wind speed of the stream, and drive the northward displacement and acceleration of the subtropical high-wind stream (Figure 2). This anomalous pattern can persist until May, ultimately triggering strong cyclonic conditions north of the stream and anticyclonic circulation to the south (Figure 3). The anticyclone prevents cold air masses penetrating from the north and blocks delivery of water vapor from the south, resulting in reduced precipitation over the LLHC. The abnormal downdraft over LLHC (Figure S4) serves to reinforce the atmospheric stability and aggravate the rainfall deficit. This teleconnection is caused primarily by altered meridional temperature gradients arising from seasonal snow cover, with secondary circulations resulting from the enhancement and displacement of the westerly high-wind stream.

### 2.3 Impacts of the cooling effect of snow cover in the linear baroclinic model

We conducted simulations using the linear baroclinic model (LBM) to further verify the dynamics and distal
climatic impacts of snow cover. Details of the model configuration are provided in the supplemental material. As shown in Figure S2, the modeled temperature forcing anomalies near the surface (1,000 hPa) are as extreme as −4 K during JFM, but decrease sharply until the 800 hPa level before stabilizing at approximately −1.5 K between 800 and 300 hPa. Above this level, the abnormal temperature forcing declines sharply, becoming negligible above 200 hPa. Our simulation also indicates that the strength of the temperature forcing varies by season. For instance, the temperature anomaly is greatest during DJF and JFM; vertical profiles for these periods exhibit the strongest cold forcing near the surface and warm forcing in the upper atmosphere. Declining snow cover in early spring causes mitigatory forcing in the vertical profile, while the sharp decline in snow cover during...
spring corresponds to the smallest temperature forcing (maximum 1 K) for the period MAM. This seasonal evolution of the vertical temperature profile aligns with the output of our composite analysis described earlier.

The added temperature forcing component to the LBM results in zonal wind anomalies, in line with the thermal wind theory (Figure S2). Specifically, an easterly wind anomaly develops at 60°N, located between the cold anomaly at 50°N and the warm anomaly between 70° and 80°N. Likewise, we observed the development of a westerly wind anomaly at 40°N, to the south of the cold anomaly forcing. This anomalous wind pattern exhibits a symmetrical barotropic structure extending vertically from 700 to 50 hPa. The intensity of the simulated wind anomaly peaks at the 200 hPa level, in accord with our composite analysis (Figure 1), while the direction of the wind anomalies reverses near the surface, resulting in westerly wind anomalies at 60°N and easterly wind anomalies near 40°N. Such reversals reflect abnormally shallow high pressure near the surface (Figure S3).

The cold temperature forcing profile causes sinking immediately above the surface, resulting in a shallow high-pressure system in the overlying level that inverts wind directions in the lower atmosphere. This idealized structure of zonal wind anomalies is a consequence of the simplified dynamical equations used in the LBM. Nonetheless, our simulations successfully reproduced the zonal wind anomaly patterns and validated the westerly wind anomalies at 80°N that are caused by the meridional temperature gradient north of the forcing region. We also note that the intensity of the simulated westerly stream varies seasonally, being strongest in winter–late winter and weakening in the spring, similar to our observed profiles and composite results.

The simulated zonal wind anomalies align with the geopotential height anomalies shown in Figure 2e–h. In response to the negative temperature forcing profile at 50°N, a low-pressure anomaly develops in the atmosphere above the forcing region. Correspondingly, anomalous high pressure north of the forcing region is associated with regionally warmer temperatures. The geopotential height anomalies exhibit barotropic-like vertical structures, expanding from the surface to the top of the atmosphere, while the abnormally high geopotential
Simulated geopotential height responses share the same temporal variation with the temperature forcing profiles; geopotential height signals are greatest in DJF, JFM, and FMA and weakest in MAM. As in the temperature forcing profiles, the strongest signal is observed in JFM. The temperature forcing profile produces a cold temperature forcing below the 200 hPa level. By altering the meridional temperature gradient, LBM produced a low-pressure anomaly over the forcing region that propagates eastward to impact downstream regions including Mongolia and China (Figure S4h). Correspondingly, a high-pressure anomaly located north of the forcing region also propagates eastward, impacting northern Siberia.

Similar to the findings of previous studies (Ye, 2019) and our composite analysis (Section 2), we observed an (N)AO-like pattern in our simulations of geopotential height. However, this regional pattern is ephemeral and does not persist until the spring. The cyclonic and anticyclonic circulations over Mongolia and the LLHC, respectively, are both reproduced in our LBM simulations. We note that the simulated zonal wind anomalies persist until spring, creating cyclonic conditions over Mongolia that weaken the Mongolia High and thus prevent cold air from flowing south. The concurrent anticyclone over the LLHC (Figure S4h) blocks the arrival of moisture to the region. Minor differences in circulation between our LBM simulations and composite analysis probably reflect the inability of the simplified LBM model to reproduce specific wavelets. Therefore, we do not anticipate close alignment between the modeled vertical velocity patterns and those derived from our composite analysis (Figure S4).

### 3 DISCUSSION AND CONCLUSIONS

In this investigation of the effects of Eurasian snow cover on atmospheric circulation downstream, we observed a significant relationship between snow cover and precipitation over the LLHC. As demonstrated by the leading SVD mode, increased snow cover in Europe and western Asia is associated with reduced precipitation over the LLHC in May (the first month of the regional wet season), and vice versa. This relationship accounts >59% of the variability in LLHC precipitation during that month. We propose that the broad distribution of Eurasian snow cover generates cold temperatures near the surface due to negative impact on the upward longwave radiation flux. In previous studies, snow cover was thought to influence the overlying atmosphere primarily through the snow-albedo and snow-hydrology effects, which we have represented in our assessment using surface-upward shortwave radiation and surface latent heat flux. Nevertheless, we observed that both effects serve to cool the atmosphere above, which in turn influences atmospheric circulations downwind.

Acknowledging that previous research has established the roles of the Indian and East Asia Summer Monsoons in LLHC precipitation (Cao et al., 2012, 2014) and the impact of ENSO on that snow–monsoon connection (Yang, 1996), we found that the snow–precipitation link is in fact independent from ENSO. Instead, we propose a different teleconnection by which Eurasian snow cover influences LLHC rainfall in May, involving the development of opposing zonal wind anomalies in response to snow-driven shifts in the meridional temperature gradient. These anomalous zonal winds in turn serve to accelerate and displace the subtropical high-wind stream. We suggest that this abnormal circulation spreads east via the westerly stream and persists throughout the boreal winter–spring season, whereupon it is accompanied by cold cyclonic conditions over Mongolia and anticyclonic circulation over the LLHC in springtime. This circulation pattern, combined with regionally sinking air (Figure 4), contributes to the overall deficit of May rainfall over the LLHC.

Recognizing the simplified nature and limited resolution of the LBM model, our simulations nonetheless reproduce the observed patterns of snow-driven atmospheric circulation, thereby confirming the dynamic linkage described earlier. Our LBM simulations also validated the seesaw pattern in the geopotential height difference between the middle and polar latitudes, namely the (N)AO-like pattern that occurs over Europe due to the temperature forcing profile in central Europe (Ye, 2019).

We note that this study focuses solely on the relationship between Eurasian snow cover and LLHC precipitation on seasonal-to-interannual timescales; the linkage on multidecadal timescales was not investigated. Nonetheless, this study affords new insight into snow–atmosphere coupling by establishing how snow cover affects the high-wind regions of Asia via alteration of meridional springtime temperature gradients, and how this process impacts regional precipitation. We also highlighted this snow–precipitation linkage as a plausible climatic impact of snow cover. Other mechanisms include the generation of a Rossby wavelet (Wang et al., 2020), which might also help explain various distal impacts of Eurasian snow cover.

Future works should focus on estimating the extent to which snow cover affects downstream atmospheric
circulation, the performance of snow–atmosphere coupling on different timescales, and how will this snow–precipitation link might change in response to anthropogenic global warming. In addition, higher-resolution numerical models with better representation of atmospheric processes, both horizontally and vertically, are required to simulate the full climatic ramifications of snow cover and to identify the specific mechanisms of the snow–precipitation linkage.

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CONFLICT OF INTEREST
The authors declare no conflicts of interest.

AUTHOR CONTRIBUTIONS
Anqi Ma: Data curation; formal analysis; investigation; software; writing - original draft. Jingchuan Zhao: Formal analysis; investigation; writing - original draft. Lei Cai: Formal analysis; investigation; validation. Zeyu Dong: Data curation; formal analysis; validation; visualization. Ruowen Yang: Conceptualization; investigation; writing-review & editing.

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