Impacts of Spatiotemporal Anomalies of Tibetan Plateau Snow Cover on Summer Precipitation in Eastern China

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ABSTRACT

Tibetan Plateau (TP) snow cover undergoes significant temporal and spatial variations during the winter and spring months. This study investigates the relationship between the spatiotemporal distribution of winter–spring snow cover (SC) over the TP and summer precipitation in eastern China (EC) using the singular value decomposition (SVD) method. Four simulation experiments are designed to validate the results of SVD analysis. Both observations and simulations show that heavier snow cover in the southern TP leads to more rainfall in the Yangtze River basin and northeastern China, and less precipitation in southern China, whereas heavier snow cover in the northern TP results in enhanced rainfall in southeastern and northern China and weakened precipitation in the Yangtze River basin. The linkage is attributed to anomalous westerly winds in the upper troposphere at around 200 hPa and to changes of the southern branch of westerlies at 500 hPa on the south side of the TP, which are caused by lasting diabatic heat anomalies over the TP. The shifts in position of the westerly jet at the exit region and negative anomalies of geopotential height at 500 hPa further result in anomalous anticyclone over the East China Sea and the corresponding 850-hPa water vapor convergence and influence the anomalous summer precipitation belt in EC.

1. Introduction

As an external forcing, snow plays an important role in the global radiation balance (Shukla and Mooley 1987; Sankar-Rao et al. 1996; Walland and Simmonds 1996) and atmosphere–land interaction (Yeh et al. 1983; Vernekar et al. 1995). It also has strong effects on the energy budget, hydrologic processes, and atmospheric circulation anomalies, which are regarded as sensitive indicators of climate change (Blanford 1884; Yeh et al. 1983; Bamzai and Shukla 1999). Changes of snow in the Northern Hemisphere (Déry and Brown 2007; Brown and Mote 2009; McCabe and Wolock 2010; Brown and Robinson 2011; Shi and Wang 2015) have greatly influenced the regional and global climate (Cohen and Entekhabi 1999; Kumar and Yang 2003) and caused remote precipitation anomalies (Chen et al. 2008; Ding and Gao 2015).

Numerous studies have demonstrated that changes in the spatiotemporal distribution of snow cover over Eurasia in response to a changing climate are complex (Brown 2000; Ye and Bao 2001; Liu and Yanai 2002; Brown and Mote 2009; McCabe and Wolock 2010; Brown and Robinson 2011; Brutel-Vuilmet et al. 2013). Snow cover over Eurasia strongly influences Northern Hemisphere climate via snow–hydrological effects and albedo effects on land surface energy budgets and moisture storage (Barnett et al. 1989; Douville and Royer 1996; Fasullo 2004; Peings and Douville 2010). However, the different spatiotemporal allocations of Eurasian snow cover have different impacts on later climate. For example, an analysis of snow cover data from the National Oceanic and Atmospheric Administration (NOAA) (Bamzai and Shukla 1999) showed that only winter snow cover over western Eurasia has a close negative relationship with Indian summer monsoon precipitation, and that snow cover over the Tibetan Plateau (TP) has no significant relationship.

Because of the high topography of the TP in the Northern Hemisphere midlatitudes, at the westerly

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upstream area of the East Asian monsoon region, its dynamic and thermodynamic effects on weather and climate in East Asia have received much attention (Ye and Gao 1979; Wu et al. 2007; Duan and Wu 2008; Wu et al. 2009; Liu et al. 2012; Wu et al. 2012a,b; Boos and Kuang 2010, 2013; Duan et al. 2013). In addition, there are both a large amount of frozen ground and extensive snow cover over the TP, which may impact weather and climate. Wang et al. (2003) suggested that hydrothermal changes caused by the freeze–thaw process are an external heat forcing that influences East Asian climate and is closely related to summer rain belts over China. Although a part of the Eurasian snow cover, the effects of snow cover over the TP on the establishment of the East Asian monsoon (Zwiers 1993; Wu et al. 2012a,b), the variability of rain belts (Wang et al. 2000; Wu and Qian 2003; Wang et al. 2008), and the associated anomaly of large-scale circulation (Wang et al. 2003) are distinguished from those in other regions of Eurasia (Yasunari et al. 1991; Zhang et al. 2004). Furthermore, snow cover on the TP has undergone significant spatio-temporal changes in past decades (Qin et al. 2006; Wang et al. 2009; Bo et al. 2014). The range and thickness of snow in the Himalayas is revealed to be negatively correlated with summer precipitation in northeastern India (Blanford 1884; Walker 1910). The establishment of the Asian summer monsoon depends on the influence of the TP through its dynamic and thermodynamic effects on atmospheric circulation (Hahn and Manabe 1975; Ose 1996; Zhang et al. 2004; Liang et al. 2005). A significantly stable relationship between TP snow cover and summer precipitation over the Yangtze River basin and southern China has been clearly defined through diagnosis and numerical analysis (Chen 1997, 1998, 2001; Chen and Wu 2000; Zheng et al. 2000; Qin et al. 2006; Ding et al. 2010). Recently, Qian et al. (2011) suggested that snowpack pollution would result in the increase of TP diabatic heating and enhancement of the East Asian monsoon, leading to anomalously wet, dry, and slightly wet patterns over southern China, the Yangtze River basin, and northern China, respectively. The winter–spring snow cover over the TP influences spring and pre-rainy season precipitation over northwestern China and eastern China (EC) (Xu and Li 1994; Wang et al. 2000). In addition, changes in snow cover over the TP have a close relationship with the extent of the mei-yu (Wu and Qian 2003). However, studies of the impacts of TP snow cover have mainly focused on the effect of snow cover over the whole TP. Qian et al. (2003) investigated the relationship between winter–spring snow over the TP and spring–summer precipitation in China using singular value decomposition (SVD) analysis and model simulation. They suggested that a positive correlation exists between winter snow over the southern TP and spring precipitation in southern China, and a negative correlation in the Huaihe River basin; there is also a positive correlation between winter snow over northern TP and summer precipitation in the Yangtze–Huaihe River basin. Although they noticed the different impacts of winter–spring snow in different regions of TP on spring and summer precipitation in EC, they still emphasized the importance of the impacts of snow over the whole TP and anomalies of snow in eastern TP in their conclusions. The possible physical mechanism of the model result was explained by impacts of TP snow anomalies on the establishment of Asian summer monsoon. Zhao and Qian (2007) investigated the impacts of anomalies of ground temperature in spring over different regions of the TP, further suggesting that ground temperature anomalies in May over northeastern and northwestern TP have different influences on summer atmospheric circulation over East Asia. The clouds, moisture recycling, and precipitation over TP have distinct spatial differences (Wang and Guo 2012; Guo and Wang 2014; Wang and Wang 2015; Wang et al. 2012, 2015a,b). Observations show the prominent spatiotemporal distribution characteristics of TP snow cover (Bo et al. 2014) and diabatic heating (Zhao and Qian 2007) of the TP. Thus, the impacts of snow cover on climate in EC vary over different parts of the TP.

Snow cover has been used as an important indicator of summer precipitation in EC in summer seasonal operational forecasting (Chen 1997, 1998, 2001; Chen and Wu 2000). However, many uncertainties still remain. As the majority of meteorological observational stations are located in the eastern and southeastern TP, and in valley regions, snow cover data are incomplete. The uneven distribution of observational stations and the various ways in which parameters such as snow cover and snow depth are measured result in inconsistencies in the collected data. For example, Zheng et al. (2000) and Qian et al.(2003), using observational snow data, reported that TP snow cover in winter has a stronger relationship with summer precipitation than spring precipitation, and that it is the dominant influence on precipitation over southern China and the middle and lower reaches of the Yangtze River. Chen et al. (2000), using NOAA/NESDIS satellite snow data, suggested that TP winter snow is positively related with precipitation over the middle and lower reaches of the Yangtze River, especially the northern Jiangnan region. Years in which TP snow is anomalously heavier or lighter than normal are different in different snow datasets (Chen et al. 2000). The relationship between the spatiotemporal distributions of TP winter–spring snow cover and summer precipitation in EC is still uncertain. Therefore, the uncertainties in the
relationship between snow cover over the TP and summer precipitation in EC and the underlying physical mechanism are worth exploring.

Variations in the spatial distribution of snow cover over the TP will lead to correspondingly varying exchanges of energy and water between land and atmosphere, thereby affecting diabatic heating, and further changes the TP thermal forcing to the atmosphere (Li et al. 2001; Zhang and Tao 2001; Zhu et al. 2007). The anomalies of TP thermal forcing could greatly influence the subtropical westerly and easterly on the northern and southern sides of the TP, respectively. Numerical model results demonstrate that the TP thermal forcing accelerates the higher-latitude westerly on its west and lower-latitude easterly on its east in summer, forming anticyclonic circulation on both sides (Liu et al. 2007a,b; Wu et al. 2014). The different heating pattern induces changes in the meridional temperature and pressure gradients across the two sides of the TP, thus changing the westerly and easterly streams over the region, as required by the geostrophic balance (Xu et al. 2002; Wu et al. 2014). Furthermore, the thermal forcings on the northern and southern branches of the westerlies caused by snow cover anomalies in the northern and southern regions of the TP should differ from each other. Observations show that summer precipitation in EC is connected with the activity of the 500-hPa trough in the Bay of Bengal, which extends from the southern branch of the westerlies caused by TP blocking (Ye and Gao 1979; Benn and Owen 1998; Liu and Yin 2001; Wu and Qian 2003), and is also connected with the activity of the subtropical jet.

This paper aims to answer the following two questions: How do different distributions of snow cover on the TP influence summer precipitation in EC? What is the large-scale mechanism driving this relationship? To answer these questions, we must deepen our understanding of the forcing of snow cover on the TP, which will improve the reliability of this indicator for predicting summer precipitation in EC.

The remainder of this paper is organized as follows. The experimental data and methodology are described in section 2. The associations between different patterns of TP snow cover and summer precipitation in EC are analyzed in section 3. A series of simulation experiments conducted to validate the results from SVD analysis are described in section 4. The underlying physical mechanism of the relations revealed by the experiments is explored in section 5. Finally, the conclusions are provided in section 6.

2. Data and methodology

This study uses the daily snow-depth dataset for 1980–2009 derived from SMMR, SSM/I, and AMSR-E passive microwave remote sensing data (horizontal resolution of 0.25° × 0.25°) obtained from the Cold and Arid Regions Environmental and Engineering Research Institute at the Chinese Academy of Sciences, Lanzhou, China (Che et al. 2008). Because the main TP snow season generally begins in October, the duration of each data record runs from October in one year to May in the following year. To reduce the temporal uncertainty of snow cover, the duration of snow is further subdivided into six periods: late autumn [October–December (OND)], early winter [November–January (NDJ)], winter [December–February (DJF)], late winter [January–March (JFM)], early spring [February–April (FMA)], and spring [March–May (MAM)]. For each time segment, monthly snow depth is defined as the maximum snow depth. Geographically, the TP covers the region 25°–40°N, 75°–105°E, where the altitude is above 2000 m. Daily precipitation data from the same time periods were collected from 756 observation stations of the National Climate Center of the Chinese Meteorological Administration, and monthly precipitation is obtained by averaging daily data. For comparison with model simulations, we used National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis atmospheric circulation data (Kalnay et al. 1996).

Singular value decomposition (SVD) is employed to investigate the collocation patterns of correlation between TP snow cover and summer precipitation in EC. SVD is an extension of empirical orthogonal function (EOF) analysis, which is designed to explain as much of the mean-squared temporal covariance between two time-dependent fields as possible. Here the two fields are the TP winter–spring snow depth (“left” field) and summer precipitation in EC (“right” field). The two time-dependent fields are each expressed as the linear combination of singular vectors. A pair of correlations is produced (i.e., the so-called left or right SVD field). For example, the pattern of the SVD first mode for winter–spring snow depth (left) is the projection of the first principal component of the EOF for summer precipitation (right). Similarly, the pattern of the SVD first mode for summer precipitation is the projection of the first principal component of EOF for winter–spring snow depth. SVD is a combination of EOF and correlative analysis (Prohaska 1976; Lanzante 1984; Bretherton et al. 1992; Wallace et al. 1992).

The statistical significance of SVD modes was tested using the Monte Carlo method; that is, the SVD left and right fields are replaced with random matrix data sampled from a Gaussian distribution, and for each pair of random matrix data SVD is performed 100 times with different random inputs; the explained variance of each
pattern for each SVD analysis is then ranked from low to high. If the explained variance for SVD analysis from actual data (snow cover and precipitation) is greater than the tenth ranked random case, this pattern is significant at the 90% confidence level (Preisendorfer and Barnett 1977; Wallace et al. 1992).

To validate the patterns obtained from SVD, simulated experiments are performed with the Community Earth System Model (CESM, version 1.2.0). CESM was developed by NCAR with fully dynamic atmosphere, land, river runoff, sea ice, and ocean components. The atmosphere component is the Community Atmosphere Model, version 4 (CAM4), and the land component is the Community Land Model, version 4 (CLM4), both at 2.5° × 1.9° (longitude by latitude) resolution. We conducted one control experiment and three sensitivity experiments to investigate the impacts of anomalies in the spatiotemporal distribution of snow cover over the TP on summer precipitation in EC.

### 3. Relationship between snow cover over the TP in winter–spring and summer precipitation in EC

To investigate the spatiotemporally collocational relationship between snow cover over the TP and summer precipitation in EC, we perform SVD analysis for normalized snow depth and precipitation data. Table 1 lists the percentage variance contribution (%) of the first three modes to the total variance in different periods from late autumn to spring. On average, the cumulative square covariance contribution of the first three modes is approximately 49%, with most of the variance explained by the first mode. The small difference in explained variance between the third and fourth modes indicates that these two modes are not independent of each other according to the criterion suggested by North et al. (1982); thus, we only analyze the first three modes that are statistically significant at the 90% confidence level according to the Monte Carlo experiments.

Table 1. Variances accounted by the first three SVD modes (%) for six different time periods.

| Time Period     | First modal | Second modal | Third modal | Total   |
|-----------------|-------------|--------------|-------------|---------|
| Late autumn     | 24.43       | 15.79        | 10.75       | 50.97   |
| Early winter    | 22.83       | 16.92        | 11.07       | 50.29   |
| Winter          | 23.51       | 19.12        | 10.45       | 53.08   |
| Late winter     | 23.75       | 20.90        | 10.00       | 54.65   |
| Early spring    | 25.85       | 20.13        | 9.43        | 55.41   |
| Spring          | 28.28       | 9.60         | 7.88        | 45.76   |

As shown in Fig. 1b, panels 1–4, one basic pattern of the relationship between snow cover over the TP and summer precipitation in EC from late autumn to later winter is found from the second SVD mode: heavier (lighter) snow cover in the western TP [hereafter referred to as the western pattern (WP)] leads to increased (decreased) precipitation in southern and northeastern China. In early spring, snow cover anomalies over the TP become smaller than in winter, and corresponding summer precipitation anomalies appear in the Yangtze River basin and northern China. Anomalies of snow cover almost disappear in spring.

The third SVD mode (Fig. 1c, panels 1–4) reflects another basic pattern describing the relationship between snow cover over the TP and summer precipitation over EC. It reveals that heavier (lighter) snow cover in the northern TP from late autumn to early spring...
Fig. 1. Typical SVD spatial patterns of heterogeneous correlations between the snow cover over TP and the summer precipitation in eastern China (EC) for six periods from autumn to spring in the subsequent year for first three leading SVD modes: (a) the first leading mode [(a1), (a2) OND; (a3), (a4) NDJ], (b) the second leading mode [(b1), (b2) DJF; (b3), (b4) JFM], and (c) the third leading mode [(c1), (c2) JFM; (c3), (c4) FMA].
results in excessive (deficient) summer precipitation in the Yangtze–Huaihe River basin and deficient (excessive) precipitation in northeastern China.

The results of the SVD analysis (Fig. 1)—that there are correlations between the spatiotemporal distributions of snow cover anomalies over the TP from winter to spring and the summer precipitation in EC—can be summarized as the three main patterns shown in Fig. 2.

4. Numerical simulations

a. Experimental design

To validate the three patterns from the SVD (Fig. 2), a control experiment (CTL) and three sensitivity experiments (SEs) are conducted as follows.

1) CTL is a control experiment without modifying the albedo calculated in the model.
2) SE1 has increasing albedo over the southern TP from November to March of the next year (corresponding to the first SVD mode, as in the shaded region in Fig. 3a).
3) SE2 has increasing albedo in the western TP from late autumn to early spring of the next year (corresponding to the second SVD mode, as in the shaded region in Fig. 3b).
4) SE3 has increasing albedo in the northern TP from late autumn to early spring of the next year (corresponding to the third SVD mode, as in the shaded region in Fig. 3c).

Considering the effect of snow cover anomalies primarily on surface albedo, the value of the surface albedo is modified to 0.9 in regions with maximum interannual variability of snow cover, which is obtained from the patterns of snow cover from SVD analysis, to represent the changes of snow cover over the TP. The surface albedo is calculated by CLM based on weighted combinations of “soil” and snow albedos (0.5–0.9), and soil albedos (0.04–0.5) that vary with soil color class based on observed MODIS local solar noon surface albedo values (Oleson et al. 2010). Therefore, the value of the albedo in CTL is slightly smaller than 0.9 with the existence of snow. The SE albedo is 0.9, implying that snow always remains. Note that each experiment was run for 20 model years, the...
first 10 years of which are for model spinup, whereas the results of the last 10 years are used for analysis. The last 10 years of simulations represent a 10-member ensemble. While the interannual variation of albedo in CTL would affect the difference in albedo between SE and CTL, there is still a positive anomaly of albedo.

b. Results

Figure 3 shows the spatial distributions of summer precipitation differences in EC from the simulations, which are generally consistent with the main modes of SVD. In the case of increasing snow cover (albedo) in the southern TP (Fig. 3a), model simulations reproduce the spatial pattern of precipitation of the leading SVD mode (Fig. 2d); i.e., precipitation increases over the Yangtze–Huaihe River basin, whereas it decreases over southern China (Fig. 3d). With the snow cover increasing in the western TP, the anomalies of precipitation are positive in the northern part of southern China and negative in the lower reaches of the Yangtze River and southwestern China (Fig. 3e), which agrees with the distribution of summer precipitation from the second SVD mode (Fig. 2e). In line with the third SVD mode (Fig. 2f), heavier snow cover in the northern TP results in excessive summer precipitation in the northern Yangtze–Huaihe River region and coastal regions of northeastern and southern China, with deficient summer precipitation in the lower reaches of the Yangtze River and southwestern China (Fig. 3f). Figure 3 implies that CESM simulations can reproduce the relationships between snow cover on the TP and summer precipitation in EC detected from observations by using SVD.

There are some differences between the distributions of precipitation anomalies from CESM simulations and SVD results, such as the negative anomalies of precipitation over southeastern China in SE2 from CESM that are the inverse of the pattern of the SVD second mode, and the positive anomalies of precipitation over northeastern China in SE3 that are the inverse of the...
pattern of the SVD third mode. These differences between SVD results and model simulations may be due to the effect of snow albedo on the radiation budget, which leads to surface cooling over the TP in winter–spring, and which is only considered in the sensitivity experiments. The inadequacies of the sensitivity experiments may reflect fewer hydrologic effects from snowmelt.

5. Mechanism of the impact of snow cover over the TP on summer precipitation in EC

Comparing Figs. 2 and 3, it is evident that the model performed well in reproducing the results obtained from SVD. To explore the mechanism underlying the impact of snow cover over TP on summer precipitation in EC, Fig. 4 shows the monthly averaged difference of TP surface net longwave radiation (NLR), surface sensible heat flux (SH), surface latent heat flux (LH), and outgoing longwave radiation (OLR) at the top of the atmosphere (W m$^{-2}$) for each month from March to August, and the JJA mean derived from (a) SE1 minus CTL, (b) SE2 minus CTL, and (c) SE3 minus CTL. The differences over the regions shaded with light (dark) color are statistically significant at 80% (90%) level under the Student’s $t$ test. Positive (negative) values are shaded in red (blue).
radiation absorbed by the ground increases in May, the cooling effect of snow weakens. Meanwhile, the increase in surface temperature over the TP results in enhanced soil thawing and snow melting, leading to an increase in soil liquid water and the intensification of LH. Nonetheless, the decrease of NLR and SH results in the weakening of TP surface diabatic heating in summer. With heavier snow in the western TP, the LH decreases, NLR and SH increase, and TP surface diabatic heating also weakens in summer. As snow changes the energy exchange between the land surface and atmosphere, less heat is transmitted to the atmosphere and the anomaly of OLR is negative (Fig. 5), which implies that the atmosphere above heavy snow cover regions is colder because of melting snow. With increased snow cover in the southern TP, the south remains cold while the northern region is warmer in summer. When snow cover in the northern TP is heavier, the whole TP is relatively cold, especially in the north. This occurs because the cold in the north results in a south–north temperature gradient and cold air in the north prevents or diminishes the warming in the south. It is worth noting that melting of heavy snow in the southern TP in June occurs earlier than the heavier snowfall in the north in July. In other words, the cooling effect of heavier snow cover in the northern TP lasts longer than in the south. Figure 5 also shows that the effects of snow cover can persist into summer, particularly until June, which is the early period of the onset of the East Asian summer monsoon.

A notable characteristic of the general circulation near the TP during the spring transition season is that the vertical cell changes from the Hadley cell to the monsoon cell. To examine the changes in vertical structure with regional snow cover changes, Fig. 6 shows a latitude–height cross section of the difference of meridional temperature gradient, zonal wind, and vertical meridional circulation along 80°–100°E between CTL and SEs. When comparing these results with the climatological mean (1981–2010) from NCEP–NCAR reanalysis, CESM reproduced well the structure of the

![Fig. 5](image-url)

**Fig. 5.** (a)–(c) The differences of snow depth (cm). (d)–(f) The differences of soil liquid water (blue solid line; mm $^3$ mm $^{-3}$), soil ice (blue dashed line; mm $^3$ mm $^{-3}$), and precipitation (black line; mm day $^{-1}$). (g)–(i) The differences of NLR (W m $^{-2}$), SH (W m $^{-2}$), LH (W m $^{-2}$) at the surface, and OLR at the top of atmosphere (W m $^{-2}$) for each month from March to August. Results are for (a),(d),(g) SE1 minus CTL, (b),(e),(h) SE2 minus CTL, and (c),(f),(i) SE3 minus CTL. The differences are averaged over the regions where the albedo is modified.
temperature gradient and the unique monsoonal cell over the TP. The direction of positive temperature gradient is defined as the increase of temperature from the south to north. Because of the snow cooling effect, the heavier snow cover over the southern TP (Fig. 3a) leads to relatively cold near-surface to lower-atmospheric temperatures over the southern TP, which also causes the heating to weaken. This corresponds to weaker ascent over the southern TP, and the decrease of temperature gradient on the southern side of TP weakens the easterly flow on the southern side of TP, and the increase of temperature gradient on northern side of the TP weakens the westerly flow on the northern side of TP in summer (Fig. 6c). With heavier snow cover in the western TP (Fig. 3b), which causes heating above the northern region to weaken dramatically, the anomalous low temperature in the western TP weakens the ascent of the northern branch of the monsoon cell and also restricts ascent over the main TP region. Colder anomalies of air at the northern side of the TP are accompanied by the decrease of temperature gradient on the northern side of TP, which accelerates the westerly flow (Fig. 6d). The temperature in the southern TP is almost unchanged, since the albedo in this region is unchanged and consequently the cooling effect is not distinct. The increasing LH from snowmelt is small in Fig. 4b. When heavier snow cover occurs in the northern TP, the lower atmosphere over the entire TP is
anomalously cold, while the atmospheric heating above the main body of the TP is enhanced, forced by surface cooling and accompanying the development of upward motion. The anomalously warm air at high latitudes is accompanied by the increase of temperature gradient on the northern side of the TP, which decelerates the westerly flow (Fig. 6e).

Figure 6 shows that with heavier snow in the southern, northern, and western TP during winter–spring, the atmospheric heating and circulation structures above the TP in summer are different. With heavier snow cover than the normal in the southern TP, the anomalous cold at whole levels restrains the rising motion over the southern TP. Changes in snow cover over the northern TP would weaken near-surface heating over the TP, forcing the enhanced atmospheric heating above the TP, which generally enhances the monsoon cell.

Changes of TP diabatic heating caused by snow cover in different regions (southern, western, and northern TP) affect downstream weather systems through atmospheric waves. A distinct feature is the change in westerly and easterly flow on the north and south sides of the TP, which results in anomalies of weather and climate in downstream regions. To further explore the impacts of changes in diabatic heating over the TP on quasi-stationary planetary waves and subtropical westerly flow in East Asia, Fig. 7 shows a zonal average cross section of June–August (JJA)-mean Eliassen–Palm (EP) flux and the divergence of EP flux. On average (Fig. 7a), the convergence at 150 hPa near 25°N establishes an easterly flow on the south side of the TP and there is a northward shift of the westerly flow, whereas the divergence of the EP flux at 250 hPa near 40°N results in enhanced westerly flow on the north side of the TP in summer. Thus, TP diabatic heating accelerates the westerly flow on the north side of the TP and the easterly on the southern side in the upper troposphere. The results from CTL (Fig. 7b) reproduce this pattern well, but the divergence of EP flux in the mid-to-high troposphere above the TP is slightly larger, especially on the 250-hPa level over the northern TP. A waveguide extends downward and southward from the upper–middle troposphere at midlatitudes (around 40°N) to the lower and southern atmosphere (around 25°N), and the southward branch only reaches the southern end of the TP in summer. This represents the effects of momentum and heat meridional transport on the zonal mean flow. The diabatic heating anomaly in the southern TP caused by heavier snow cover in the southern TP (Fig. 7c) leads to weakening divergence of EP flux at 250 hPa and 40°N, which is accompanied by decelerated subtropical westerly over the northern TP. When snow cover in the western TP is more extensive, anomalies in EP flux divergence in the mid-to-high troposphere are weaker than in SE1, which implies that the downstream effects are also weaker than in SE1. With heavier snow cover in the northern TP, the EP flux divergence decreases over the southern TP and increases over the northern TP. This situation favors the propagation of waves toward the south and enhanced westerly flow. The anomalies of EP flux in the case of larger snow cover distribution over different regions of the TP (Fig. 7) show the impacts on the zonal mean flow (especially the subtropical westerly jet) in the troposphere of diabatic heating caused by snow cover anomalies over the TP.

To examine the interaction between diabatic heating over the TP and westerly and easterly flow on the two sides of the TP, Fig. 8 shows time–latitude cross sections of zonal kinetic energy \[ u^2 \] and meridional momentum flux \[ \overline{u^2 v^2} \] (where \( u^* \) and \( v^* \) represent the mean zonal deviation of zonal wind \( u \) and meridional wind \( v \), respectively) (Hoskins and Pearce 1983) associated with stationary waves at 200 hPa in the climatological mean and SEs. On average (Figs. 8a,b), the maximum center of zonal kinetic energy in winter (November–February) is located around 30°N, which is also a zone of interaction between the TP and higher-latitude regions. The meridional momentum flux near 30°N is positive, indicating northward dynamic and thermodynamic forcing, but negative on the southern side of the TP (20°N), which appears to be the forcing of the TP onto lower latitudes. In summer, there are two maxima of meridional kinetic energy, one at 20°N, which corresponds to easterly flow over the southern TP, and another at 40°N, which represents northward movement and enhancement of westerly flow. These distribution characteristics of the generation zone and the meridional momentum flux relate to changes in the mean flow.

The differences in zonal kinetic energy and meridional momentum fluxes between CTL and SE are shown in Figs. 8c–d. With heavier snow cover in the southern TP (SE1), the zonal kinetic energy shows a marked increase on the southern side of the TP, and a slight decrease on the northern side of the TP in winter. In summer (JJA), the zonal kinetic energy decreases on both sides of the TP, especially on the north, which corresponds to the reduced EP flux divergence over the northern TP. The northward transport of meridional momentum flux also diminishes (Fig. 8c). With heavier than normal snow in the western TP, the zonal kinetic energy decreases over the southern TP and increases over the northern TP from November to December. Zonal kinetic energy decreases over the entire TP in the subsequent spring. In summer, the zonal kinetic energy shows a marked increase on both sides of the TP, and the northward transport of meridional momentum flux is also enhanced.
When snow cover on the northern TP is heavier, the changing pattern of zonal kinetic energy is similar to that in the case of heavier snow on the southern TP in winter, but anomalies are weaker and only persist from January to February. In summer, the changes in zonal kinetic energy and meridional momentum flux are similar to those in the case of heavier snow cover in the west. While zonal kinetic energy only changes in the northern TP, the changes are not as great as in SE2 and they persist for only a short period (from June to July) in the subsequent summer.

**Figure 8** shows that the changes in dynamic and thermodynamic forcing caused by snow cover anomalies not only occur in the TP, but also affect the interaction between the TP and the atmosphere in subsequent summers. The impacts of heavier snow cover in the southern and western TP on dynamic and thermodynamic forcing of the TP wave on the southern and northern sides of the TP regime are remarkable in later summer. In other words, the anomalies of diabatic heating in different regions of the TP in winter–spring lead to different downstream effects in subsequent summers.
To further reveal the impacts of regional snow cover anomalies on circulation in subsequent summers, Fig. 9 shows the differences of JJA mean vorticity, horizontal wind at 850 and 200 hPa, and 500-hPa potential temperature between SE and CTL. Heavier snow cover over the southern TP results in cold anomalies in the south and warm anomalies in the north at 500 hPa (Fig. 9b). Correspondingly, an anomalously cold cyclone and positive vorticity form over the southern TP at 200 hPa (Fig. 9a). In the downstream region, the anomalous cyclone is projected over the Great Khingan Range (50°N, 115°E) and anomalously negative vorticity is seen over the Korea Peninsula (40°N, 130°E), accompanied by an anomalous cyclone and anomalously positive vorticity near the East China Sea (25°N, 127.5°E). The anomalies of 500-hPa geopotential height (GH) are negative, and the western Pacific subtropical high (WPSH) is slightly weaker than normal (Fig. 10b). Correspondingly, an anomalous anticyclone also exists over the East China Sea at 850 hPa (Fig. 9c). Nevertheless, with heavier snow in the northern TP, the atmosphere on the north side of the TP is cold at 500 hPa, with the anomalous cyclone and positive anomalies of vorticity at 200 hPa. The differences with respect to the case of heavier snow in the southern TP are the anomalous cyclone and negative vorticity over northeastern
China, which is also accompanied by an anomalous cyclone near the East China Sea. The prominently negative anomalies of GH represent weakening of the WPSH at 500 hPa. At low level, an anomalous anticyclone is produced at 850 hPa. The cooling effect of heavier snow cover on the western TP is seen clearly at the northwest side of the TP and adjacent northern region, which leads to the anomalous cyclone and positive anomalies of vorticity in the upper atmosphere (200 hPa) over the northern TP. Accordingly, a strong anomalous anticyclone occurs over the East China Sea at 850 hPa. The remarkable differences in atmospheric circulation anomalies in EC are the anomalous cyclone from SE1 and anticyclone from SE2 and SE3 near the Sea of Japan at 200 hPa, which further lead to the northward shift of the anomalous anticyclone over the East China Sea at 850 hPa from SE2 and SE3.

A comparison of wind circulation anomalies in summer caused by heavier snow cover in different regions of the TP reveals that the impacts on the atmospheric circulation in EC result from change to the subtropical westerly jet. There are no obvious effects on weather systems over the southern TP. Wang et al. (2008) reported that the impacts of the TP on the low-level atmosphere arise through a distinct Rossby wave train moving along the monsoon westerlies, as resolved in a linearized AGCM.

To clearly show interactions between regional diabatic heating on the TP and weather systems in East Asia, Fig. 10 shows changes in 200-hPa zonal winds, 500-hPa GH, and transport of water vapor at 850 hPa. The definition of westerly jet is $u > 0 \text{ m s}^{-1}$ and $|V| \geq 30 \text{ m s}^{-1}$ (Schiemann et al. 2009). With heavier snow cover on the southern TP, the westerly flow at 200 hPa is weakened over the northwestern TP and enhanced near the Sea of Japan (Fig. 10a), and the position of the westerly jet moves southward at the East Asia jet exit region. These changes result from the anomalous cyclone and anticyclone over the TP and northeastern China, respectively (Fig. 9a). With heavier snow cover on the western TP, the westerly flow at 200 hPa is enhanced over the northern TP, while a remarkable weakening of westerlies occurs over the Sea of Japan, which results in the northward shift of the westerly jet exit region. When snow cover over the northern TP is

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**Fig. 9.** The differences of JJA mean vorticity (color shading; $10^{-5}$ s$^{-1}$) at (top) 200, (middle) 500, and (bottom) 850 hPa from (a)–(c) SE1 minus CTL, (d)–(f) SE2 minus CTL, and (g)–(i) SE3 minus CTL; grid cells with values significant at 90% level under the Student’s $t$ test are marked by black dots. The vectors in each panel represent the differences of horizontal wind (m s$^{-1}$); green vectors represent significance at 90% level under the Student’s $t$ test. The contours represent the differences of the potential temperature (°C); red solid and blue dashed contours represent the positive and negative values of potential temperature difference significant at 90% level under the Student’s $t$ test, respectively.
heavier, the westerly flow is weakened over the northern TP, and the changes in zonal winds over East Asia are similar to those resulting from heavier snow cover in western TP. The impacts of snow cover in different regions of the TP on the westerly flow are distinct from the inverse changes in position of westerly jet at exit region, which are accompanied by different changes in the westerly field in the upper troposphere.

Changes in GH (Figs. 10b,e,h) show that heavier snow cover in different regions of the TP consistently lead to negative anomalies of GH over the TP. With heavier snow cover on the southern TP, the GH is slightly weakened in EC, which causes the slight weakening of the WPSH. With increased snow cover in the western and northern TP, the weakening of the WPSH is pronounced.

With an increase in regional snow cover over the TP, water vapor transport and areas of convergence correspond well to the distribution of precipitation anomalies in Fig. 3. The strength and location of anomalous anticyclones are obviously different with heavier snow cover in different regions of the TP. Heavier snow cover over the southern TP causes an anomalous anticyclone over the East China Sea (30°N, 125°E) at 850 hPa, which results in more water vapor being transported to the Yangtze River region and less water vapor in southern China; thus, precipitation is excessive in the Yangtze River region and deficient in southern China. When snow cover on the western and northern TP is heavy, the anomalous anticyclone shifts northward over Japan (34°N, 130°E), implying increased water vapor. The distribution of differences in JJA mean vapor flux divergence (Fig. 10c) agrees with the distribution of summer precipitation anomalies. This implies that heavier winter–spring snow cover in the northern TP leads to the northward shift of the distribution of summer precipitation anomalies in EC.

6. Conclusions

The observed relationship between regional snow cover over the TP and summer precipitation in EC was
investigated, and three main spatial collocations were found from the first three SVD modes. With heavier snow cover on the southern TP, the subsequent summer precipitation increases in the upper reaches of the Yangtze River and decreases in southern China. Heavier snow cover on the western TP leads to excessive summer precipitation in northeastern and southern China. Increased snow cover on the northern TP is often accompanied by enhanced summer precipitation over southeastern and northern China and reduced summer precipitation over the Yangtze River basin. Simulations using CESM generally reproduce the main relational patterns.

The mechanism by which increased snow cover in different regions of the TP affects summer precipitation in EC is also explored. The results of numerical experiments show that with heavier snow cover on the southern TP, the cooling effects result in a decrease of air temperature gradient and negative anomalies of the westerly jet at the northern side of the TP at 200 hPa, which cause a southward shift of the westerly jet exit region and an anomalous cyclone. With heavier snow cover in the western and northern TP, the westerly jet moves northward at the exit region, accompanied by an anomalous anticyclone.

Diabatic heating anomalies caused by snow cover anomalies in different regions of the TP also lead to anomalous westerlies at 200 hPa along the whole TP, which influence the convergence zone where the southern branch of the westerly flow and the northern branch of the westerly flow converge in EC, as well as the westerly flow on the southern side of the TP at 500 hPa. Along with anomalies of westerly jet, the negative anomalies of GH in eastern China influence the WPSH activity at 500 hPa. Compared with previous studies that emphasized the role of an anomalous anticyclone over the Philippines (Wang et al. 2008; Wu et al. 2005), this paper reveals the mechanism of the impact of snow cover changes in different regions of the TP on summer precipitation in EC.

Corresponding to the heavier snow cover in the southern TP, the anomalous anticyclone at 850 hPa over the East China Sea (30°N, 125°E) will transport more water vapor to the Yangtze River valley and less to northern China, which results in excessive precipitation in the upper Yangtze River and deficient precipitation toward the south. The difference between snow cover anomalies over the southern TP and over the northern TP are that the latter results in a stronger anomalous anticyclone, located farther north (over Japan; 34°N, 130°E), leading to more water vapor convergence over northern China and a northward shift of summer precipitation in EC. The atmospheric circulation anomalies in EC result from the anomalous cold in the north (caused by heavier snow cover on the southern TP), which gives rise to the enhanced subtropical westerly jet in the exit region, while heavier snow cover in the northern TP weakens the subtropical westerly jet. These results are summarized in Fig. 11.

This study investigated the relationship between snow cover anomalies in different regions of the TP and summer precipitation in EC on large scales. However,
the underlying physical mechanism needs further investigation. The present results provide new evidence that snow cover in different regions of the TP can be used as a predictor signal for precipitation in EC, as well as in investigating the development of the summer monsoon in East Asia.

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