A Bridging Role of Winter Snow over Northern China and Southern Mongolia in Linking the East Asian Winter and Summer Monsoons

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ABSTRACT: Eurasian snow, one of the most important factors that influence the Asian monsoons, has long been viewed as a useful predictor for seasonal monsoon prediction. In this study, observations and model simulations are used to demonstrate a bridging role of the winter snow anomaly over northern China and southern Mongolia (NCSM) in the relationship between the East Asian winter monsoon (EAWM) and the East Asian summer monsoon (EASM). Enhanced snow in NCSM results in local surface and tropospheric cooling, strengthening the EAWM through cold-air intrusion induced by northerly wind anomalies. In turn, the stronger EAWM provides a favorable condition for enhanced snowfall over East Asia to the south, indicating an active snow–EAWM interaction. The continental cooling could be maintained until summer due to the memory effect of snowmelt and moistening as well as the snow–monsoon interaction in the spring, causing changes in the meridional temperature gradient and associated upper-level westerlies in the summer. The interaction between the strengthened westerlies over the northern Tibetan Plateau and the topography of the plateau could lead to anomalous downstream convergence and compensating divergence to the south. Therefore, anomalous cyclonic circulation and increased rainfall occur over northeastern China and the Korean Peninsula, but anticyclonic circulation and decreased rainfall appear over the subtropical East Asia–Pacific region. Moreover, limited analysis shows that, compared to sea surface temperature feedback, the direct impact of snow anomaly on the EAWM–EASM connection seems more important.

KEYWORDS: Monsoons; Climate variability

1. Introduction

The East Asian monsoon system regulates the variations of atmospheric circulation, temperature, and precipitation over East Asia all year round. The East Asian winter monsoon (EAWM) and the East Asian summer monsoon (EASM), which are characterized by northerly wind and southerly flow respectively, are often associated with strong cooling in winter (Lau and Li 1984; Ding 1994; Chen et al. 2005; Chang et al. 2006) and heavy rainfall in summer (Tao and Chen 1987; Ding 1994; Chang et al. 2000).

The variations of EAWM are strongly affected by, or highly related to, high-latitude systems including the Siberian high, the Ural blocking, the Arctic Oscillation (AO), among others (e.g., Ding and Krishnamurti 1987; Gong et al. 2001; Wu and Wang 2002; Cheung et al. 2012; Sun et al. 2016). In fact, many of these systems such as the Siberian high, the East Asian westerly jet stream, and East Asian cold surge have been regarded as EAWM components (Chang and Lau 1982; Zhang et al. 1997; Jeong and Ho 2005; Park et al. 2010), and the EAWM influences weather and climate through these high-latitude systems. The EAWM is also interacted with tropical systems/phenomena including El Niño–Southern Oscillation (ENSO), the Madden–Julian oscillation (MJO), and tropical convection more generally (e.g., Li 1990; Wang et al. 2000; Li and Yang 2010). Recent studies further showed that the EAWM could be measured by a northern mode and a southern

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mode, and its northern mode was related to high-latitude phenomena including Arctic sea ice variability but the southern mode was influenced more strongly by ENSO (Wang et al. 2010; Li and Yang 2017).

Previous studies have documented a generally negative relationship between the EAWM and the EASM: a weak EAWM is often followed by enhanced summer monsoon rainfall over East Asia, especially central China, and vice versa (Chen et al. 2000, 2013). Given their importance for regional economy and society, variations of the EASM and associated droughts, flooding, and other disasters have long been a subject of substantial research interest. Establishing and understanding the relationship between EAWM and EASM would be potentially important for operational EASM prediction. Chen et al. (2013) demonstrated a role of ENSO in bridging the winter–summer monsoon connection. The winter monsoon variability related to ENSO is largely responsible for the observed EAWM–EASM relationship, while the non-ENSO part of winter monsoon variability shows no evident association with the following EASM. However, ENSO may not be the only factor that affects this monsoon connection; instead, the variation of Indian Ocean sea surface temperature (SST) can also play a role in the EAWM–EASM relationship (Wang and Wu 2012).

Land surface processes have long been considered an important factor for monsoon variability and in this context a number of previous studies have focused on snow conditions (e.g., Blanford 1884; Hahn and Shukla 1976; Parthasarathy and Yang 1995; Yang and Lau 1998; Notaro and Zarrin 2011; Turner and Slingo 2011; Wu et al. 2012; Halder and Dirmeyer 2017). Variations of Eurasian cold-season snow influence surface radiation, heat exchange between land and the atmosphere, land temperature, and soil moisture through snow-albedo and snowmelt-related hydrological effects, causing significant changes in weather and climate (Barnett et al. 1988, 1989; Cohen and Rind 1991; Cohen 1994; Cohen and Entekhabi 1999, 2001; Cohen et al. 2001; Henderson et al. 2018). It has been found that Eurasian snow exhibits a strong association with the variations of EAWM (e.g., Jhun and Lee 2004; Wang et al. 2009; Wang et al. 2010; Luo and Wang 2019). More Eurasian snow leads to surface cooling, a cold winter, and a strong EAWM (Watanabe and Nitta 1998; Jhun and Lee 2004; Wang et al. 2009, 2010). According to Jhun and Lee (2004), a lack of autumn snow over East Asia can induce local surface warming and corresponding rising motion, and then a compensating downward motion over the North Pacific, resulting in weakening of the Aleutian low and the EAWM.

Compared to the impact on EAWM, the impact of Eurasian snow on East Asian summer rainfall is more complex (e.g., Yang and Xu 1994; Liu and Yanai 2002; Yim et al. 2010; Zhang et al. 2017). Yang and Xu (1994) demonstrated that Eurasian winter snow cover was related negatively with the summer rainfall in the middle and lower reaches of the Yangtze River, but positively with the rainfall in North and South China. Its relationship with the summer rainfall averaged for all of China was insignificant. Yim et al. (2010) also depicted nonuniform features of the snow–EASM relationship: two distinct patterns of spring Eurasian snow-cover anomaly existed but the dipole pattern, rather than the uniform one, was more closely

connected with the summer rainfall in East Asia. Recently, Zhang et al. (2017) demonstrated that Eurasian spring snowmelt exhibited a dominant mode with an east–west dipole structure; that is, less snowmelt over the Siberia and more snowmelt around Lake Baikal. This anomalous snowmelt pattern and associated soil moisture anomalies could trigger an anomalous Rossby wave, which may then interact with the climatological quasi-stationary wave, resulting in more summer rainfall in Northeast China and less rainfall in most of the south.

The relationships of Eurasian snow with both EAWM and EASM reviewed above raise several questions: Does the Eurasian snow play a role in bridging the EAWM and EASM variations? If yes, where is the possible impacting snow anomaly located? What are the seasonal and interannual features of the snow–monsoon relationship and their associated physical processes? Both observational analysis and model experiments will be conducted to address these questions in this study.

The data and model used in this study, as well as the model experiment design, are introduced in section 2. Section 3 illustrates the characteristics of observed relationships between winter snow cover, EAWM, and EASM. In section 4, the responses of land conditions, atmospheric circulation, and rainfall to winter snow perturbation in a coupled climate model are discussed. The physical processes and possible mechanisms involved are also demonstrated. A discussion about the role of SST variations induced by snow perturbation in the EAWM–EASM relationship is given in section 5. Conclusions and a further discussion are provided in section 6.

2. Data, model, and experiment design

a. Observational data

Version 4 of the Northern Hemisphere EASE-Grid 2.0 snow cover used in this study is obtained from the National Snow and Ice Data Center (<http://nsidc.org/data/>) with the original weekly data at a grid cell size of $25\text{ km} \times 25\text{ km}$, which has been converted to monthly mean with $1^\circ \times 1^\circ$ spatial resolution. These snow-cover data are derived from the Advanced Very High Resolution Radiometer (AVHRR), the Geostationary Operational Environmental Satellite (GOES) System, and other visible-band satellite data (Brodzik and Armstrong 2013). Monthly precipitation data are from the Global Precipitation Climatology Project (GPCP) version 2.3 (Adler et al. 2003), which is provided by the NOAA/OAR/ESRL PSD (<https://www.esrl.noaa.gov/psd/>) with a spatial resolution of $2.5^\circ \times 2.5^\circ$. Monthly skin temperature, soil water, and atmospheric variables are acquired from the European Centre for Medium-range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim) with a spatial resolution of $1^\circ \times 1^\circ$ (Dee et al. 2011). The period for our analysis is from January 1979 to December 2016.

b. Model

The NCAR Community Earth System Model version 1.2.2 (CESM 1.2.2) is a state-of-the-art Earth system model, with interactive components of the atmosphere, land, ocean, and sea ice (Hurrell et al. 2013; Neale et al. 2013). The atmospheric component CAM4 possesses a horizontal resolution of $1.9^\circ \times 2.5^\circ$ and 26 vertical levels. The resolution of the ocean

component POP2 is approximately 1° with 60 vertical levels. All model experiments are implemented with the B2000 and F2000 component sets, which are for coupled model and atmospheric model experiments respectively, with carbon dioxide, aerosol, solar forcing, and ozone concentration fixed at their values in year 2000.

c. Experimental design

A fully coupled integration of 250 years is conducted to spin up the model, and the following 100-yr integration is taken as the control experiment by the coupled model (CTRL_cpl). To understand the impact of winter snow anomaly on the East Asian monsoon and the role of snow-induced SST variations, several sensitivity experiments are conducted. The first run is a snow perturbation experiment using the coupled model (Snow_cpl). It consists of 1-yr integrations restarted from every 1 October of CTRL_cpl but with a constant additional snowfall rate of 1.1 mm day^{-1} in northern China and southern Mongolia (NCSM; $35^\circ\text{--}47.5^\circ\text{N}$, $70^\circ\text{--}125^\circ\text{E}$) from December to the following February, according to the observed pattern of anomalous winter snow cover related to EASM variability.

Before separating the direct impact of snow from the modulation of SST variations on the change in EASM rainfall, we first verify that the main features in Snow_cpl can be reproduced by the atmospheric model. To do so, two atmospheric model experiments are performed. One is the control experiment (CTRL_atm), driven by the 100-yr SST output of CTRL_cpl, and the other one is the sensitivity experiment (Snow_pertbst), which is driven by the 100-yr SST of Snow_cpl with the same snowfall perturbation as in Snow_cpl. The aim here is to keep the key parameters including the SST and snow perturbations the same as in Snow_cpl, even though the two-way coupling between the atmosphere and the ocean is now disabled. As reviewed in the introduction, it can be reasonably assumed that the snow perturbation and the SST variation induced by the snow perturbation are the most important boundary forcings in influencing the East Asian climate.

Two additional sensitivity experiments by the atmospheric model are conducted to quantify the effects of snow and SST, respectively. One is Snow_ctrlsst, driven by the SST of CTRL_cpl and the same snow perturbation introduced above, revealing the direct impact of the snow. The other one is Nosnow_pertbst, driven by the SST of Snow_cpl but without snow perturbation, reflecting the role of SST variations induced by the snow forcing in the snow–monsoon relationship. The outputs of 100 integrations for all experiments are analyzed and the differences between the sensitivity and control experiments are compared to understand the impacts of the snow anomaly and the associated SST variations on the East Asian monsoon climate.

3. Observational features of winter snow and EASM–EASM relationship

a. Relationship between winter snow anomaly and EASM variability

To investigate the snow–EASM relationship, the EASM index defined by Wang et al. (2009) as the difference in winter

FIG. 1. (a) Standardized time series of EASM index with its linear trend and the linearly regressed signal of ENSO removed. (b) Correlation between winter snow cover and the EASM index. White dots indicate that the correlation coefficients are above the 90% confidence level. The topography of the Tibetan Plateau above 1500 m is marked with the dashed green contour. The red box indicates the location of snow-cover anomaly most strongly associated with the winter monsoon. (c) Standardized time series of snow-cover index related to EASM variability, with its linear trend and the linearly regressed signal of ENSO removed.

sea level pressure between the Siberian high and the Aleutian low is analyzed. This large pressure contrast between the two systems, with strong northerlies along the coastal region of East Asia, is one of the major characteristics of the EASM (Ding 1994). According to the definition of this index, which is highly correlated with that defined by Gong et al. (2001) based on the sea level pressure over Siberia, the EASM is strong when the index is positive, and vice versa. Figure 1a provides the standardized time series of EASM index with its linear trend and the linearly regressed signal of ENSO removed, which shows large interannual variability. To determine the locations of snow-cover anomalies that are most strongly associated with EASM variability, a map of correlation between winter snow cover and EASM index is given in Fig. 1b. It can be seen that the winter snow-cover anomaly in the NCSM

FIG. 2. Composite differences in winter (a) surface temperature (shading; unit: $^{\circ}\text{C}$), (b) sea level pressure (shading; unit: Pa) and horizontal wind at 850 hPa (vectors; unit: m s^{-1}), and (c) 200-hPa zonal wind (shading; unit: m s^{-1}) and wind vectors (vectors; unit: m s^{-1}) between more and less snow-cover years. White dots and blue vectors indicate the values above the 90% confidence level. Purple contours in (c) show winter climatology of 200-hPa westerlies. The topography of Tibetan Plateau above 1500 m is marked with the dashed green contour.

region is closely related to monsoon variability: a stronger EAWM is associated with more snow cover. Therefore, a snow-cover index is defined as the area average of winter snow cover over NCSM (see the red box in Fig. 1b) and the corresponding standardized time series is shown in Fig. 1c (the coefficient of correlation between the EAWM index and the snow-cover index is 0.39, which exceeds the 95% confidence level). Years with enhanced and reduced snow covers are then selected based on the values of snow-cover index greater than 0.9 or less than -0.9 standard deviation. The values of ± 0.9 instead of ± 1 are selected to ensure more samples for subsequent composite analyses of the possible impacts of winter snow anomaly on the monsoon. The change in winter snow water equivalent associated with the snow-cover index can reach up to 15 mm in eastern NCSM, indicating a generally positive relationship between snow cover and snow amount (figure not shown). The hydrological effect of snow depth on

FIG. 3. As in Fig. 2, but for (a) spring surface temperature (shading; unit: $^{\circ}\text{C}$), and (b) summer precipitation (shading; unit: mm day^{-1}) and 850-hPa horizontal winds (vectors; unit: m s^{-1}). The letter C indicates anomalous low-level cyclonic circulation and A indicates anomalous anticyclonic circulation.

the climate in following months is critical, since the albedo feedback of snow does not have such long persistence although it is also important in the cold seasons (e.g., Barnett et al. 1989; Yasunari et al. 1991; Vernekar et al. 1995).

The changes in winter surface temperature and atmospheric circulation related to the snow-cover index are given in Fig. 2. Corresponding to a greater snow-cover anomaly in the NCSM region, surface temperature decreases over the midlatitude Eurasian continent, especially over East Asia and Mongolia (Fig. 2a). Moreover, the Eurasian continental high and the nearby oceanic low intensify, with anomalous northerly wind along the coast of eastern Eurasia (Fig. 2b). At the upper troposphere, the extratropical westerlies become stronger, especially on the southern side of jet entrance and the northern side of jet exit (Fig. 2c) where atmospheric baroclinic instability is strong. All these results, depicting a close NCSM snow–EAWM connection, characterize the major features of a strong EAWM (Ding and Krishnamurti 1987; Yang et al. 2002; Jhun and Lee 2004; Chang et al. 2006; Li and Yang 2010; Wang et al. 2010). It should also be pointed out that these monsoon features are often accompanied by cold-season hazardous weather events such as severe storms (Wen et al. 2009; Zhou et al. 2011) and cold surges (Chang and Lau 1980; Boyle and Chen 1987; Chan and Li 2004; Park et al. 2011; Li et al. 2017).

FIG. 4. Climatology of (a) winter surface temperature (shadings, unit: $^{\circ}\text{C}$) and sea level pressure (contours; unit: Pa), (b) spring surface temperature (shading; unit: $^{\circ}\text{C}$) and 850-hPa winds (vectors; unit: m s^{-1}), and (c) summer precipitation (shading; unit: mm day^{-1}) and 850-hPa winds (vectors; unit: m s^{-1}) in ERA-Interim. (d)–(f) As in (a)–(c), but for the climatology in the CTRL_cpl. The topography above 1500 m is marked with the dashed green contour.

b. Relationship between winter snow anomaly and variability of EASM rainfall

Figure 3a shows that the colder surface temperature anomaly over East Asia following more NCSM winter snow and a stronger EAWM is maintained persistently into the spring season, associated with an increase in spring snow water equivalent in the NCSM region (figure not shown). Associated with the change in upper-level zonal wind and its interaction with the topography due to the change in meridional temperature gradient (figure not shown), anomalous convergence and cyclonic circulation occur over northeastern China and the Mongolian region and anomalous divergence and anticyclonic circulation appear over the subtropical East Asia and western Pacific region. Correspondingly, decreased summer rainfall appears over subtropical East Asia and the Pacific, with increased rainfall to the north (Fig. 3b). Therefore, Fig. 3 reveals

an apparent relationship between the NCSM winter snow and spring surface temperature and summer rainfall over East Asia. However, these observed associations do not necessarily imply causation. Model experiments are needed to investigate the possible casual effect of NCSM snow on the East Asian climate in the subsequent seasons and the associated physical processes and dynamical mechanisms.

4. Impact of winter snow on East Asian monsoon in a coupled climate model

Figure 4 depicts the climatological features of global circulation and surface temperature in both reanalysis data (left panels) and the model control run CTRL_cpl (right panels). Overall, CTRL_cpl captures the observed patterns shown in Figs. 4a–c, with a warm bias over much of the world in boreal winter and spring. In particular, the bias in the surface

FIG. 5. Differences in winter (a) snowfall (shading; unit: mm day^{-1}), (b) snow depth (shading; unit: mm), and (c) surface temperature (shading; unit: $^{\circ}\text{C}$) between Snow_cpl and CTRL_cpl. Red box indicates the domain of snow perturbation. White dots indicate the values above the 90% confidence level. The topography of Tibetan Plateau above 1500 m is marked with a dashed green line.

temperature over East Asia is small in winter but a warm bias is apparent in spring. The model generally overestimates summer precipitation over the southern Tibetan Plateau (TP) and the tropical oceans but underestimates precipitation over the northern Bay of Bengal, the equatorial eastern Indian Ocean, the Korean Peninsula, and Central America. Overall, East Asia is among the regions better simulated by the model.

a. Response of winter atmospheric circulation to snow perturbation

To understand the impact of winter snow on East Asian monsoon, a control experiment CTRL_cpl with the coupled CESM is conducted and a corresponding sensitivity experiment Snow_cpl is set up by adding a snowfall perturbation of 1.1 mm day^{-1} to every grid in the NCSM region (Fig. 5a). Compared with that in CTRL_cpl, the snow depth in Snow_cpl

FIG. 6. As in Fig. 5, but for winter (a) sea level pressure (shading; unit: Pa) and 850-hPa horizontal winds (vectors; unit: m s^{-1}), (b) 500-hPa geopotential height (shading; unit: m) and horizontal winds (vectors; unit: m s^{-1}), and (c) 200-hPa zonal wind (shading; unit: m s^{-1}) and wind vectors (vectors; unit: m s^{-1}). In (c), the purple contours outline the climatological mean of 200-hPa zonal wind in CTRL_cpl.

increases in a range of 150–300 mm (Fig. 5b). Corresponding to the anomalous snow, winter surface temperature decreases not only in NCSM but also in subtropical East Asia, indicating a strong southward cold-air intrusion (Fig. 5c) that is favorable for the formation of snowfall. It can be seen that the snowfall and snow depth in subtropical East Asia, which is to the south and thus outside the region with added snow, also increase (Figs. 5a,b), indicating an active snow–EAWM interaction. That is, cold-air intrudes southward into subtropical East Asia when more snow occurs in the NCSM, favoring snow formation in subtropical East Asia. This interactive process is consistent with the “snow–EAWM feedback mechanism” proposed by Luo and Wang (2019), in which cold-air advection from the high latitudes is enhanced as more snow and intensified high pressure appear in the Mongolian Plateau, favoring snow formation and accumulation in Mongolia and East Asia, which could in turn strengthen the Mongolian high.

In response to the snowfall perturbation, the continental surface high and oceanic low intensify with anomalous northerly wind along the coastal region of East Asia in winter (Fig. 6a), indicating a strong EAWM. The modeled feature is similar to the observed pattern shown in Fig. 2b, in spite of a

FIG. 7. As in Fig. 5, but for spring (a) soil liquid water and ice in the top 10 cm of soil (shading; unit: kg m^{-2}), (b) latent heat flux (shading; unit: W m^{-2}), and (c) surface temperature (shading; unit: $^{\circ}\text{C}$).

southward shift of the simulated pattern. In the middle troposphere, an anomalous “negative–positive–negative–positive” wave train pattern forms over the extratropical Northern Hemisphere, particularly with a deepening and slightly westward-shifting East Asian trough (Fig. 6b). The figure shows an atmospheric baroclinic structure in the Asian continent but equivalent barotropic structure in other regions. In the upper troposphere, extratropical westerlies generally strengthen, especially over East Asia and the western Pacific (Fig. 6c). The above features, including continental cooling, an intensified continental high and oceanic low, and a strengthened East Asian trough and upper-level westerlies, match the main characteristics of a strong EAWM and resemble the previous observed patterns shown in Fig. 2.

b. Changes in spring land conditions

It is important to see that the significant wintertime signals can persist to the following spring season. As shown in Figs. 7a and 7b, soil moistening and the corresponding increase in

FIG. 8. As in Fig. 5, but for summer (a) tropospheric temperature (shading; unit: $^{\circ}\text{C}$) and (b) 200-hPa zonal wind (shading; unit: m s^{-1}).

surface latent heat flux occur in the NCSM region due to snowmelt. Surface cooling could be maintained due to the land heat loss from snow melting and the increase in surface evaporation (Fig. 7c). Meanwhile, the spring snow depth in Snow_cpl is still larger than that in CTRL_cpl (figure not shown), which also contributes to the local surface cooling in NCSM. It is known that the land–ocean thermal contrast in spring begins to transform from a “colder land–warmer ocean” structure to its opposite structure of “warmer land–colder ocean,” favoring the development of summer monsoon accompanied by water vapor transport from the oceans. In short, the above features indicate that the evolution and intensity of summer monsoon could be affected by changes in land surface conditions. It is further expected that the EASM, which is an important component of the Asian summer monsoon system and is located to the south of NCSM, would change significantly if spring land conditions respond strongly to the NCSM winter snow anomaly. The physical processes and possible mechanisms involved in the snow–EASM relationship are further explored below.

c. Changes in summer atmospheric circulation and rainfall

As the boundary and tropospheric cooling persists (Fig. 8a), the upper-level zonal wind exhibits evident variations (Fig. 8b) through the change in the meridional temperature gradient. The westerly jet stream acts as a waveguide, which dynamically

FIG. 9. As in Fig. 5, but for summer (a) 850-hPa divergence (shading; unit: $\times 10^6 \text{ s}^{-1}$), and (b) precipitation (shading; unit: mm day^{-1}) and horizontal winds at 850 hPa (vectors; unit: m s^{-1}).

influences the East Asian monsoon climate (Yang et al. 2002; Chowdary et al. 2019), especially the mei-yu-baiu rainband (Sampe and Xie 2010), through eddy-mean flow interaction and atmospheric teleconnection patterns. A hypothesis has been proposed by previous studies (Molnar et al. 2010; Chiang et al. 2015, 2017; Kong and Chiang 2020) to explain the subseasonal–seasonal evolution of East Asian rainband in

warm seasons. Following the northward shift of the East Asian subtropical jet stream, the rainband in East Asia moves northward through the interaction between the upper-level westerlies and the topography of TP, which could lead to downstream surface convergence due to quasigeostrophic lifting. Similar features based on the above mechanism also emerge from our model experiments with snow forcing. According to Fig. 9a, the intensification of the upper tropospheric westerlies over northern TP is accompanied by anomalous surface convergence in northeastern Asia, downstream of the anomalous westerlies. Anomalous divergence related to the compensating circulation occurs to the south, especially over subtropical East Asia and the western Pacific. Therefore, decreased rainfall occurs over the subtropical East Asia–Pacific region during summer (Fig. 9b). However, compared with the observed rainfall change shown in Fig. 3b, the center of decreased rainfall over the subtropical Asian–Pacific region shifts southwestward slightly and the rainfall increase over northeastern Asia is smaller in Snow_cpl. The model–observation discrepancy could be due to the model’s imperfect simulations of the mean state, the various kinds of coupling and feedback processes, and other factors. Nevertheless, Snow_cpl captures the overall pattern of the observed change in East Asian summer rainfall. It should be pointed out that the results shown above are from the average of 100-yr integrations and those from the first and second 50-yr integrations both show a large resemblance, indicating the statistical robustness of the features presented.

To summarize (see the schematic diagram shown in Fig. 10), the winter snow anomaly in the NCSM region actively interacts with the EAWM by inducing continental cooling, intensified high pressure, and associated cold-air intrusion, which in turn provide a favorable condition for snow formation and accumulation in the EAWM region to the south. In the following spring, boundary and tropospheric cooling could be maintained because of the heat loss from snowmelt and the increase in surface evaporation, inducing changes in meridional temperature gradient and the upper-level westerlies. Anomalous

FIG. 10. Schematic diagram of the role of NCSM winter snow in the EAWM–EASM connection.

FIG. 11. As in Fig. 5, but for (a) winter, (b) spring, and (c) summer surface temperature (shading; unit: $^{\circ}\text{C}$).

westerlies and their interaction with the topography of TP could lead to downstream surface convergence due to quasi-geostrophic lifting and then compensating divergence to the south. Therefore, summer rainfall increases over northeastern Asia and decreases over the subtropical East Asian–Pacific region.

5. Direct effect of winter snow versus the effect of SST variations induced by snow anomaly on monsoon

Two more experiments, Snow_pertbsst and CTRL_atm, are conducted by using the atmospheric component of CESM (see detailed description in section 2c) to understand whether the snow plus SST forcing can reproduce the results from Snow_cpl. The CTRL_atm is a control experiment and the Snow_pertbsst is an experiment in which both snow perturbation and the SST change due to the snow perturbation as seen in the

control coupled model experiment (Fig. 11) are included. Over the Pacific, surface temperature does not show any apparent change during winter when snow perturbation is added (Fig. 11a), and SST signals become more apparent in the subsequent spring with cooling in the western-central Pacific and slight warming in the northeastern and equatorial eastern Pacific (Fig. 11b). The spring pattern evolves to be stronger in summer, especially the warming over the equatorial eastern Pacific, indicating an El Niño-like distribution (Fig. 11c). Furthermore, another sensitivity experiment Snow_ctrlsst, in which only snow perturbation is included without these SST changes, is performed to understand the direct impact of snow perturbation.

Figure 12 shows the differences in winter sea level pressure, 850-hPa winds, and 500-hPa geopotential height and winds between Snow_pertbsst and CTRL_atm, and between Snow_ctrlsst and CTRL_atm, respectively. Experiment Snow_pertbsst generally captures the main features shown in Snow_cpl, which is the snow perturbation experiment with the coupled model (shown in Figs. 6a,b), especially the signals in the Eurasian continent during winter (see Figs. 12a,b). However, compared with those in Snow_cpl, the North Pacific low is weaker and the high over northeastern North America is stronger in Snow_pertbsst, indicating the potential effect of two-way air–sea coupling. Furthermore, the patterns of winter circulation variations in Snow_ctrlsst resemble those in Snow_pertbsst, indicating that the direct effect of snow makes a dominant contribution to the changes in East Asian winter climate (Figs. 12c,d). For spring land conditions, the changes in soil water and temperature are highly consistent among the coupled and the two uncoupled experiments (figure not shown), indicating the importance of both local and remote impacts of the NCSM winter snow perturbation for the variability of East Asian cold-season climate.

Compared with Snow_cpl, the tropospheric cooling and the corresponding change in upper-level westerlies are well reproduced by the two uncoupled experiments in spite of a slight northward pattern shift (Fig. 13). To a certain extent, the signals in Snow_ctrlsst are weaker than those in Snow_pertbsst, implying a possible strengthening effect of SST variations on the change in East Asian summer climate. The patterns of summer low-level circulation and rainfall shown in Fig. 9b are generally reproduced by Snow_pertbsst (Fig. 14a), in which the increased rainfall over northeastern China and the Korean Peninsula is more evident. However, the pattern of decreased rainfall over the subtropical East Asian–Pacific region exhibits a more zonal distribution, compared with that shown in Fig. 9b. Nevertheless, experiment Snow_ctrlsst could explain a large portion of the change in East Asian summer rainfall in Snow_pertbsst (Fig. 14b).

To identify the role of snow-induced SST variations on the EASM, an additional atmospheric sensitivity experiment Nosnow_pertbsst is performed with the snow-induced SST variations only and without snow perturbation (see details in section 2c). Figure 14c, which shows the differences in summer rainfall and 850-hPa winds between Nosnow_pertbsst and CTRL_atm, presents a slight decrease in rainfall in the subtropical western Pacific around 35°N but without any evident signal in East Asia. This result indicates that the snow-induced SST variations contribute to the decrease in summer rainfall

FIG. 12. Differences in winter (a) sea level pressure (shading; unit: Pa) and 850-hPa horizontal winds (vectors; unit: m s^{-1}), as well as (b) 500-hPa geopotential height (shading; unit: m) and horizontal winds (vectors; unit: m s^{-1}), between Snow_pertbsst and CTRL_atm. (c),(d) As in (a) and (b), respectively, but for the differences between Snow_ctrlsst and CTRL_atm. White dots and blue vectors indicate the values of composite differences above the 90% confidence level. The topography of TP above 1500 m is marked with a dashed green line.

over the subtropical East Asian–Pacific region but this effect is relatively small compared with the direct impact of the snow.

6. Conclusions and discussion

This study is conducted to understand a previously revealed relationship between the East Asian winter and summer monsoons: a weak EAWM is followed by a strong EASM with

increased summer rainfall over central China (and vice versa) and SST variability is considered as a factor to connect the variations of the monsoons. Both observational analyses and model experiments are applied to investigate the EAWM–EASM relationship from a new perspective: the Eurasian snow bridges the variations of East Asian winter and summer monsoons. Particular effort is devoted to understanding the

FIG. 13. As in Fig. 12, but for summer (a),(c) tropospheric temperature (shading; unit: $^{\circ}\text{C}$) and (b),(d) 200-hPa zonal wind (shading; unit: m s^{-1}).

FIG. 14. Differences in summer rainfall (shading; unit: mm day^{-1}) and 850-hPa winds (vectors; unit: m s^{-1}) (a) between Snow_pertbst and CTRL_atm, (b) between Snow_ctrlsst and CTRL_atm, and (c) between Nosnow_pertbst and CTRL_atm. White dots and blue vectors indicate the values of composite differences above the 90% confidence level. The topography of TP above 1500 m is marked with a dashed green line.

direct impact of snow on the monsoons and their connection, although the indirect effect of snow-induced SST variations is also considered.

It is observed that the winter snow anomaly over the northern China–southern Mongolia region is significantly associated with both EAWM and the EASM variations. A strong EAWM is associated with more snow in NCSM, which is followed by variations of the EASM with decreased rainfall over the subtropical East Asian–Pacific region and increased rainfall over northeastern China and the Korean Peninsula. Model experiments with perturbed snowfall demonstrate the snow–monsoon relationship reasonably well and clarify some of the associated physical processes and mechanisms. Local surface

and atmospheric cooling and enhanced northerly wind with strong cold-air advection occur due to the increased NCSM snow, strengthening the EAWM as characterized by the intensified surface continental high, oceanic low, East Asian trough, and upper-level westerlies. In turn, the stronger EAWM provides a favorable condition for enhancing snow over East Asia to the south, reflecting an active snow–EAWM interaction. The cooling signal could be maintained until early summer due to the snowmelt and moistening effect as well as the interactive snow–monsoon process in spring, leading to changes in the meridional temperature gradient and upper-level westerlies. Based on past studies (Molnar et al. 2010; Chiang et al. 2015, 2017; Kong and Chiang 2020), we suggest that anomalous downstream convergence with cyclonic anomalies and its compensating divergence with anticyclonic anomalies to the south can emerge from the interaction between the strengthened westerlies over the northern TP and the underlying topography. In this way, rainfall increases over northeastern China and the Korean Peninsula but decreases over the subtropical East Asian–Pacific region. Comparing three uncoupled atmospheric model experiments (with both snow perturbation and snow-induced SST variations, with snow perturbation only, and with SST variation only) indicates that the direct impact of winter snow plays a more important role in the EAWM–snow–EASM connection than the indirect impact of SST variations, which makes a slightly positive contribution to the above relation. It should also be pointed out that compared to previous studies this investigation has applied more samples of observations and model experiments to further demonstrate the robustness of the EAWM–EASM relationship, which may provide useful information for the seasonal prediction of the East Asian summer monsoon.

It should be cautioned, however, that the current study has only addressed the snow–monsoon relationship on seasonal-to-interannual time scales, while the relationship between Eurasian snow and the Asian monsoon changes on decadal time scales (Liu and Yanai 2002; Si and Ding 2013; Zhang et al. 2019). How the Eurasian snow affects the Asian monsoon including the EAWM–EASM transition on a subseasonal time scale also remains as a subject for future investigations. Moreover, several studies have demonstrated that the TP snow influences the variability of the Asian monsoon on seasonal-to-interannual and longer time scales (Xiao and Duan 2016; Wang et al. 2018; Lu et al. 2020). It is natural to ask whether the TP snow also plays a role in the EAWM–EASM connection. In addition, in spite of a simple discussion about the impact of snow-induced SST variations in the current study, a full understanding of the relative impacts of air–sea coupling and land–air interaction on the EAWM–EASM relationship is worthy of further investigations.

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