Near-Global Atmospheric Responses to Observed Springtime Tibetan Plateau Snow Anomalies

SHIZUO LIU,^{a,b,c} QIGANG WU,^{a,b} STEVEN R. SCHROEDER,^d YONGHONG YAO,^{b,c} YANG ZHANG,^{b,c}
TONGWEN WU,^c LEI WANG,^a AND HAIBO HU^{b,c}

^a *Department of Atmospheric and Oceanic Sciences/Institute of Atmospheric Sciences, Fudan University, Shanghai, China*

^b *Institute of Global Changes, Nanjing University, Nanjing, China*

^c *Institute of Atmospheric Sciences, Nanjing University, Nanjing, China*

^d *Department of Atmospheric Sciences, Texas A&M University, College Station, Texas*

^e *National Climate Center, China Meteorological Administration, Beijing, China*

(Manuscript received 26 March 2019, in final form 4 October 2019)

ABSTRACT

Previous studies show that there are substantial influences of winter–spring Tibetan Plateau (TP) snow anomalies on the Asian summer monsoon and that autumn–winter TP heavy snow can lead to persisting hemispheric Pacific–North America-like responses. This study further investigates global atmospheric responses to realistic extensive spring TP snow anomalies using observations and ensemble transient model integrations. Model ensemble simulations are forced by satellite-derived observed March–May TP snow cover extent and snow water equivalent in years with heavy or light TP snow. Heavy spring TP snow causes simultaneous significant local surface cooling and precipitation decreases over and near the TP snow anomaly. Distant responses include weaker surface cooling over most Asian areas surrounding the TP, a weaker drying band extending east and northeast into the North Pacific Ocean, and increased precipitation in a region surrounding this drying band. Also, there is tropospheric cooling from the TP into the North Pacific and over most of North America and the North Atlantic Ocean. The TP snow anomaly induces a negative North Pacific Oscillation/western Pacific-like teleconnection response throughout the troposphere and stratosphere. Atmospheric responses also include significantly increased Pacific trade winds, a strengthened intertropical convergence zone over the equatorial Pacific Ocean, and an enhanced local Hadley circulation. This result suggests a near-global impact of the TP snow anomaly in nearly all seasons.

1. Introduction

Snow anomalies affect moisture, radiation, and the energy budget, and cause both local and distant temperature responses (Namias 1985; Cohen and Rind 1991; Groisman et al. 1994). Many studies have examined impacts of Eurasian snow cover anomalies on the hemispheric and regional atmospheric circulation. Observational and modeling studies have shown that autumn Siberian snow cover extent (SCE) anomalies can induce an Arctic Oscillation (Thompson and Wallace 1998) or North Atlantic Oscillation (Wallace and Gutzler 1981) atmospheric response in winter, via the vertical propagation of Rossby waves through tropospheric–stratospheric

coupling (e.g., Cohen and Entekhabi 1999; Gong et al. 2003; Cohen and Fletcher 2007; Fletcher et al. 2009; Smith et al. 2011). There is also some observational and modeling evidence that extensive autumn/winter Eurasian snow cover can have a downstream influence on atmospheric circulation over the Asia–Pacific–North America sector, mainly characterized by a deepened surface Aleutian low and/or a Pacific–North America (PNA)-like midtropospheric atmospheric response with a strong eastward extension toward Europe (e.g., Walsh and Ross 1988; Yasunari et al. 1991; Walland and Simmonds 1996; Clark and Serreze 2000; Orsolini and Kvamstø 2009; Wu et al. 2011). The storm tracks over the North Pacific Ocean respond strongly to Eurasian snow cover anomalies, and transient eddy forcing is important for maintaining the PNA response (Wu et al. 2011).

The TP is a small part of Eurasia, but TP snow cover variability and persistence show distinct features. First, snow cover over the Himalayan and Tibetan Plateau

Denotes content that is immediately available upon publication as open access.

Corresponding author: Qigang Wu, qigangwu@fudan.edu.cn

regions has much larger interannual variability than over other regions of Eurasia (Fasullo 2004), indicating that its associated impact may also be important at interannual time scales. Second, TP snow cover anomalies often persist from winter to spring, and sometimes even through summer over the high altitudes of the Himalayas (Pu et al. 2007; Xiao and Duan 2016). Third, as a result of an average altitude exceeding 3000 m, anomalous TP and Himalayan snow acts as an elevated net heating anomaly that plays an important role in the regional and global atmospheric circulation. Many studies have shown that net heating induced by TP/Himalayan snow-cover variability can trigger characteristic large-scale atmospheric circulation patterns, consequently leading to anomalous Asian summer monsoon rainfall (e.g., Fasullo 2004; Blanford 1884; Yanai et al. 1992; Webster et al. 1998; Wu and Qian 2003; Zhang et al. 2004; Wang et al. 2008; Turner and Slingo 2011; Xu et al. 2012; Si and Ding 2013; Xiao and Duan 2016; Senan et al. 2016; Wang et al. 2018). Anomalous (either heavy or deficient) winter-spring TP snow cover has a lingering effect as the snow melts gradually, and the resulting soil moisture anomaly then affects summer Indian and East Asian rainfall (Shukla and Mooley 1987; Bamzai and Shukla 1999; Xiao and Duan 2016).

Large positive snow cover anomalies over the TP in autumn and winter can cause a significant hemispheric PNA-like response in winter, as suggested in recent observational studies (Wu et al. 2011; Lin and Wu 2011; Mote and Kutney 2012), and demonstrated in our recent numerical study (Liu et al. 2017). In the current study, we further provide observational and modeling evidence that the global atmosphere responds significantly to spring [March–May (MAM)] TP snow anomalies. Therefore, TP snow anomalies in nearly all seasons are important sources of variability of near-global atmospheric circulation at interannual time scales. The paper is organized as follows. Section 2 describes data sources, observation analysis methods, and modeling experiments. Section 3 presents observation-based results and numerical modeling results. Section 4 summarizes conclusions.

2. Datasets and methods

The experimental procedures are almost identical to those in Liu et al. (2017), except for the specific forcing, and use the same two online satellite snow datasets, described below. The first dataset is SCE from “Northern Hemisphere EASE-Grid 2.0 Weekly Snow Cover and Sea Ice Extent, version 4” (Brodzik and Armstrong 2013) (updated 2014; available online at <https://nsidc.org/data/nsidc-0046>) covering from November 1978 to December 2015 in this study, derived from several visible-band

satellite instruments. The second dataset is snow water equivalent (SWE) from “Global Monthly EASE-Grid Snow Water Equivalent Climatology, version 1” (Armstrong et al. 2005) (available online at <https://nsidc.org/data/nsidc-0271>) covering from November 1978 to May 2007 (SWE has not been updated since that month), derived from several microwave instruments. In the EASE-Grid datasets, for each $25\text{ km} \times 25\text{ km}$ cell, SCE is a simple binary variable of the presence or absence of snow cover, and SWE is the average water content of snow cover in millimeters. Several studies have used the EASE-Grid SCE and SWE datasets to examine impacts of TP snow anomalies on the Asian summer monsoon (e.g., Zhao and Moore 2004; Xiao and Duan 2016; Wang et al. 2017). For the figures, or for model input, this study interpolates or averages both SCE and SWE data sources to a uniform grid box size of $2.8^\circ \times 2.8^\circ$, which is the same horizontal resolution of the model simulations described in modeling experiments. The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996) is used in the observational analysis, which includes global atmospheric data fields such as 500-hPa geopotential heights (Z500) and covers 1979–2015 (37 years) in this study.

In the observational analysis, maximum covariance analysis (MCA) (Czaja and Frankignoul 2002) is performed to examine lag effects of late-winter and early-spring TP SCE anomalies on 1) the MAM midlatitude circulation from central Asia eastward to North America and 2) the MAM tropical circulation. To reduce the influence of trends, a first-order polynomial is removed by least squares fit to each grid point time series. Also, we filter out the tropical Pacific (ENSO) influence using a regression against the Niño-3.4 (5°N – 5°S , 120° – 170°W) sea surface temperature (SST) anomalies of preceding months, where the regression coefficient is selected as the maximum regression coefficient within the preceding 6 months. The coupled MCA patterns associated with the first MCA mode in Figs. 3 and 4 are formed from the heterogeneous covariance map for 500-hPa geopotential height (Z500) and the homogeneous covariance map for SCE. These maps are constructed by regressing the Z500 and SCE fields onto the normalized MCA-SCE time series at each lag.

Characteristic atmospheric circulation responses to TP snow forcing strongly resemble well-known teleconnections. For comparison, an empirical orthogonal function (EOF) analysis is applied to the observed (NCEP–NCAR reanalysis) annual MAM mean Z500 from 1979 to 2015 for the North Pacific region (20° – 85°N , 120°E – 120°W , as shown in Fig. 1). The EOF analysis is area weighted by the square root of the cosine of latitude before computing the covariance matrix, but

FIG. 1. The spatial structure of the (left) first and (right) second EOFs of the seasonal MAM Z500 anomalies over the North Pacific region outlined in (a) (20° – 85° N, 120° E– 120° W) in (a),(b) the observations and (c),(d) CAM4 control run simulations. This is the same as the regression of the Z500 field onto the standardized principal component time series of the leading two EOF modes. Regression coefficients are drawn every 5 m, negative contours are dashed, and the zero contour is omitted.

all physical fields are plotted as correct values. The leading two EOFs of Z500 explain 26.0% and 23.8% of the total variance, and are readily identifiable as being approximately the PNA and western Pacific (WP) (Wallace and Gutzler 1981; Linkin and Nigam 2008) teleconnection patterns (Figs. 1a,b). The surface signature of the WP teleconnection pattern is the North Pacific Oscillation (NPO), usually defined as the second EOF in sea level pressure (SLP) in winter (Linkin and Nigam 2008). Figures 1c and 1d show that EOFs resulting from 200-yr model control run simulations are similar to corresponding observed EOFs, with spatial pattern correlations of 0.86 and 0.83 for EOF1 and EOF2, respectively, but the model EOF1 has a much stronger (positive) polar center than that in the observation.

As in Liu et al. (2017), model experiments are forced with observed heavy and light TP SCE and SWE obtained from the two EASE-Grid observational datasets specified above, with a common period from November 1978 to May 2007. Six MAM spring periods from 29 available are chosen in this study, specifically 1981, 1983, and 1998 with persistent heavy SCE and SWE over the TP and 1984, 1985, and 1992 with persistent light SCE and SWE over the TP. Figure 2 shows averages of SCE

and SWE over the TP region. The left column (labeled “pos”) is the average of heavy snow seasons of 1981, 1983 and 1998, the left-center column (“clim”) shows the climatological average from 1979 to 2007, the right-center column (“neg”) shows the average of light snow seasons of 1984, 1985 and 1992, and the right column (“pos-neg”) shows the general increases of SCE and SWE in TP heavy snow years based on the pos minus neg difference. SCE and SWE differences are greater than 20% and 25 mm, respectively from March to May over most of the TP. It should be mentioned that since the TP area is over 30° of longitude wide, typically and in these chosen years, either a heavy or light spring TP snow year usually reflects similar snow anomalies persisting since at least the preceding late autumn. Also note that SCE and SWE are taken from different datasets, and they potentially may be inconsistent. Figure 2 does not plot SCE $< 5\%$ or SWE < 5 mm, and most SCE and SWE plots have areas with concurrent SCE $> 5\%$ and SWE > 5 mm, indicating that TP SCE and SWE are generally consistent.

Model ensemble experiments are forced with observed TP SCE and SWE in the MAM periods stated above, so the model anomaly patterns and magnitudes

FIG. 2. Spring monthly (MAM, as labeled) spatial patterns of the average TP (top) SCE (%) and (bottom) SWE (mm) based on satellite observed data for (left) the average of high TP SCE and SWE years 1981, 1983, and 1998 (labeled “pos”), (left center) the climatological average from 1979 to 2007 (labeled “clim”), (right center) the average of low SCE and SWE years 1984, 1985, and 1992 (labeled “neg”), and (right) the difference pos minus neg (labeled “pos-neg”). Each panel extends from 20° to 50°N and from 60° to 110°E, and the green line outlines the TP region. The upper color bar applies to the three columns on the left, and the lower color bar applies to the right column.

should be realistic. We use the NCAR Community Atmosphere Model, version 4 (Neale et al. 2013), coupled to the Community Land Model, version 4.0 (Oleson et al. 2010). Both models have a horizontal resolution of 2.8° latitude and 2.8° longitude (T42). CAM4 is a major atmospheric model widely used in climate modeling. CLM4.0 combines realistic radiative, ecological, and hydrologic processes (Oleson et al. 2010). Both models

or their earlier versions (CAM3 and CLM3.0) have been used to investigate atmospheric responses to projected snow changes through the twenty-first century (Alexander et al. 2010), to satellite-based snow albedo changes over North America and the whole NH (Allen and Zender 2010), and to idealized snow forcing with increased albedo (Wang et al. 2017). CLM4.0 was coupled with CAM5 (updated version of CAM4) to

investigate improvements of seasonal temperature prediction when snow data are assimilated (Lin et al. 2016).

In this experiment, six ensemble simulations are performed where the prescribed forcing is TP SCE and SWE from 1981 (SNOW 81), 1983 (SNOW 83), 1984 (SNOW 84), 1985 (SNOW 85), 1992 (SNOW 92), or 1998 (SNOW 98). The ensemble simulation procedure is identical to that in our previous study (Liu et al. 2017) except for a different prescribed forcing region and different months of simulation. The SCE and SWE forcing is prescribed in the TP area (the green outlined area in each panel of Fig. 2), based on one of the six chosen years from the EASE-Grid datasets above, and model derived SCE and SWE are used everywhere else. Other surface boundary conditions, specifically SST and sea ice concentration, are climatological values from the same 200-yr control run mentioned above. Each simulation performs 100 model runs to identify atmospheric responses to extensive (SNOW81, SNOW83, and SNOW98), or limited (SNOW84, SNOW85, and SNOW92) TP snow cover, with each of the 100 runs initialized with the CAM4 atmospheric state on 1 March of different years from the control run. Each simulation uses the same 100 initializations, so each ensemble starts with identical atmospheric fields and the ensembles then diverge gradually as a result of the different imposed snow boundary conditions.

In CLM4.0, both SCE and SWE are prognostic. The parameterizations for snow and snow accumulation are based primarily on Anderson (1976) and Dai and Zeng (1997). At each time step, snow is allowed to fall and accumulate at the surface. Melting of existing snow cover releases latent heat to the atmosphere and soil moisture to the land model. However, CLM4.0 has been modified so the computed SCE and SWE in the TP area (Fig. 2) at each time step is replaced by our prescribed SCE and SWE for the specified snow season (interpolated to each model time step from monthly values). After the above modification, energy and water are not explicitly conserved in the TP area during each realization. The total climate response is the ensemble mean difference between extreme snow perturbation states, specifically the average of SNOW81, SNOW83, and SNOW98 minus the average of SNOW84, SNOW85, and SNOW92, and statistical significance is evaluated by a standard *t* test of the difference between ensemble means.

3. Results

a. Observational results

Maximum covariance analysis (MCA) (Czaja and Frankignoul 2002) is first performed to examine lagged

effects of late-winter and early-spring TP SCE anomalies on the MAM midlatitude circulation from central Asia eastward to North America. MCA identifies the patterns of two spatial fields that produce the maximum covariance. At lag = 0, the atmosphere and snow (also surface boundary SST and sea ice) connection in the MCA is dominantly characterized by the atmospheric circulation forcing on TP snow (Czaja and Frankignoul 2002; Wu et al. 2011). Therefore, the climate impacts of a TP snow anomaly on MAM atmospheric circulation can only be found by applying MCA at lag = −2 [January–March (JFM)] and −1 months [February–April (FMA)]. Specifically, MCA is applied to 1979–2015 (37 yr) of NCEP–NCAR reanalysis fields of MAM Z500 anomalies over the East Asian and North Pacific–America regions (20°–65°N, 60°E–120°W, as shown in Fig. 3b), versus monthly TP SCE anomalies averaged from the weekly EASE-Grid SCE dataset described above, for the preceding JFM (Z500 lags SCE by 2 months) and FMA (Z500 lags SCE by 1 month). The Z500 MCA pattern associated with the first MCA mode in the analyses below (Figs. 3b and 4b) is constructed by regressing the hemispheric Z500 fields from 30°S to 90°N onto the corresponding normalized MCA–SCE time series.

Figure 3 shows coupled MCA patterns associated with the first MCA mode when SCE leads by 2 months. The MCA pattern of SCE in JFM shows strongly increased SCE over the TP (Fig. 3a), and the atmospheric signal in MAM (Fig. 3b, low Z500 in the central North Pacific and high Z500 with two centers from central Siberia to Alaska) resembles the negative phase of the west Pacific (WP) teleconnection pattern (Wallace and Gutzler 1981) in the EOF analysis (Fig. 1b), with a strong high reaching 25 m over Siberia and Alaska and a low reaching −30 m over the North Pacific, related to an approximate maximum 10% TP SCE anomaly in JFM. Similar patterns are found when SCE leads by 1 month (not shown). Meanwhile, coherent negative Z500 anomalies are significant over North America and the Arctic, and positive Z500 anomalies are significant over western Europe and subtropical Pacific. The MCA patterns suggest that a consistent hemispheric-scale Z500 pattern in spring is strongly correlated with TP SCE forcing at lags of −1 or −2 months, so heavy TP snow in winter and early spring could cause a polarization of WP toward its negative phase in the midlatitudes in spring. This influence is unrelated to ENSO teleconnections and the long-term climate trend, which were removed prior to the analysis. Similar WP-like responses are also found in the stratosphere (not shown), indicating a barotropic response to TP snow anomalies in the NH troposphere and stratosphere.

The MCA is also performed between JFM or FMA TP SCE anomalies and MAM tropical Z500 (30°S–30°N,

FIG. 3. First MCA mode covariance maps based on 1979–2015 for JFM SCE over the TP region [outlined in green in (a)] against MAM Z500 over the East Asia–North Pacific area [20°–65°N, 60°E–120°W; outlined in (b)]. Note that Z500 lags SCE by 2 months. (a) Homogeneous SCE anomalies plotted over the greater TP region (total plot area is the same as in Fig. 2). (b) Heterogeneous Z500 anomalies plotted globally from 30°S to 90°N. Note that the Z500 pattern in (b) is constructed by regressing the hemispheric Z500 fields from 30°S to 90°N onto the corresponding normalized MCA-SCE time series. SCE units are percent as depicted in the color bar. Z500 units are meters, and the light-blue dashed and light-red solid contours represent -1 and 1 m, respectively. The darker blue and red contours are at intervals of 5 m, negative contours are dashed, and the zero line is omitted. Dots in (a) and shaded areas in (b) denote significant regression coefficients of SCE or Z500, respectively, on the MCA-SCE time series at the 95% confidence level based on the two-sided Student's t test.

0°–360°E). At lag = -2 months (JFM SCE vs MAM Z500), the MCA patterns in Fig. 4 suggest that increased SCE over the TP in JFM is significantly correlated with MAM positive Z500 in the tropics and negative Z500 over the North Atlantic Ocean. Large but insignificant anomalies in the midlatitudes appear in the regression due to internal atmospheric variability or coherent changes of Z500 in NH midlatitudes and in the tropics. Similar MCA patterns are found at lag = -1 month

(FMA SCE vs MAM Z500, not shown). Note that significant Z500 anomalies are mostly in the midlatitudes in Fig. 3, but in the tropics in Fig. 4. This indicates that the effect of JFM TP forcing on the MAM negative WP teleconnection is almost independent of the effect of JFM TP forcing on the MAM tropical circulation. In fact, the correlation between the MCA-SCE coefficient time series associated with the SCE patterns in Figs. 3a and 4a is only about 0.1.

FIG. 4. As in Fig. 3, but for the first MCA mode covariance maps for JFM SCE against MAM tropical Z500 (30°S–30°N, all longitudes). The darker blue and red contours in (b) are at intervals of 2 m.

FIG. 5. Monthly mean responses (MAM, as labeled; W m^{-2}) of (left) net shortwave radiation (positive downward), (left center) net longwave radiation (positive upward), (right center) sensible heat fluxes (positive upward), and (right) latent heat fluxes (positive upward). Each response is the ensemble mean “pos minus neg” difference between the high and low simulations, or the average of SNOW81, SNOW83, and SNOW98 minus the average of SNOW84, SNOW85, and SNOW92. Statistical significance of the mean difference between these two ensembles is determined using the t test (black dots indicate $>95\%$ statistical confidence level).

Overall, these observational results indicate that TP snow anomalies can induce a hemispheric atmosphere circulation response in spring. Such conclusions based on MCA results at lags -2 and -1 can be extended to the impacts of simultaneous MAM TP snow forcing since either heavy or light TP snow anomaly in late winter and early spring usually persists for several months. This will be demonstrated in numerical experiments in section 3b and the associated mechanisms will be examined in section 3c.

b. Modeled seasonal mean climate responses to TP snow forcing

Model ensemble experiments were performed as explained in section 2, forced by observed SCE and SWE anomalies from three heavy and three light TP snow years (Fig. 2), but model-generated SCE and SWE outside the TP. Analyses below discuss “pos minus neg” ensemble average differences of the heavy minus light TP snow years. The increased TP SCE and SWE forcing first affects the atmosphere regionally by changing surface energy fluxes. Due to increased albedo in heavy TP snow cover locations, the TP net surface shortwave

radiation is negative (Fig. 5), but becomes positive in May in the northeastern TP where positive SWE anomalies were small in March and April, and the snow has melted completely before May. Gradual snowmelt moistens the soil, which continues to reduce ground temperature and net longwave radiation, sensible heat and latent heat (Shukla and Mooley 1987), with magnitudes of net sensible heat fluxes generally exceeding latent heat fluxes as a result of prescribed moderately dry climatological conditions over the TP regions. Opposite effects of snow forcing on these fluxes are simulated over East China, with net surface shortwave radiation anomalies positive and much larger than other components.

Figures 6 and 7 display the ensemble mean MAM seasonal average responses to TP snow anomalies in various atmosphere variables. The surface air temperature (SAT) (Fig. 6a) is especially cold over the TP region due to negative sensible and latent heat fluxes, but is also moderately cold over many midlatitude remote land and ocean areas. Warming SAT responses are significant over regions such as eastern China, eastern Siberia and the adjacent Arctic Ocean, and the eastern North Pacific. In Figs. 6b and 6d, tropospheric circulation responses to

FIG. 6. Ensemble-mean MAM averages of the atmospheric responses of eight variables: (a) SAT, (b) SLP, (c) T500, (d) Z500, (e) T70, (f) 70-hPa geopotential heights (Z70), (g) wind pattern at 1000 hPa (UV1000), and (h) UV300. The color bar depicts temperature anomaly values [K; in (a), (c), and (e)]. Each response is the average of SNOW81, SNOW83, and SNOW98 minus the average of SNOW84, SNOW85, and SNOW92 (or “pos minus neg”). In (b), (d), and (f) (SLP, Z500, and Z70 anomalies, respectively), the light-blue dashed and light-red solid contours represent -0.05 and 0.05 hPa for SLP and -0.5 and 0.5 m for Z500 and Z70 to show the details of atmospheric responses in the tropics, and the zero contour is omitted. The darker blue and red contours (negative contours are dashed) are at intervals of 0.25 hPa for SLP and 4 m for Z500, starting with ± 0.25 hPa for SLP and

increased TP SCE and SWE clearly indicate a baroclinic vertical structure [opposite sea level pressure (SLP) and Z500 anomalies] over the TP region, and a remote equivalent barotropic structure (SLP and Z500 responses of the same sign) over East Asia and the North Pacific. Over the North Pacific, SLP shows the negative phase of the NPO (Wallace and Gutzler 1981), and the Z500 response highly resembles the WP teleconnection pattern, with pattern correlation about -0.88 in the East Asia to North Pacific region when compared with EOF2 in model control simulations. This confirms the observational results that a positive snow anomaly in the TP in spring can lead to a remote teleconnection similar to the negative phase of the WP pattern, as shown in Fig. 3b over the western North Pacific. A WP-like response is also found in the stratosphere (Fig. 6f), with the most significant negative or positive response over Asia or eastern Siberia, respectively, which suggests that dynamical mechanisms are similar between the 500- and 70-hPa levels, even though those levels are above and below the tropopause. However, air temperatures at 500 and 70 hPa (T500 and T70, respectively; Figs. 6c,e) instead show nearly opposite response patterns in the northern extratropics. The effects of horizontal temperature advection from warmer or colder regions by the anomalous flow [the wind pattern at 300 hPa (UV300) in Fig. 6h is similar at 500 and 70 hPa] play an important role in setting up the T500 and T70 patterns. The climatological temperature decreases or increases from the equator toward the North Pole in the troposphere or stratosphere, respectively. So, the WP-like cyclonic wind anomaly around 30°N and anticyclonic anomaly around 60°N in eastern Siberia account for the cold and warm T500 patterns in Fig. 6c, but generate an almost exactly opposite pattern of T70 anomalies in Fig. 6e.

Precipitation anomalies are directly related to water vapor transport/convergence and vertical motion. Significant negative precipitation responses are simulated in a band from the northwestern TP to Japan and northeastward to Alaska (Fig. 7a), induced by descending motion over the same band (Fig. 7d; increasing pressure indicates descent) associated with negative SLP and positive Z500 anomalies in Fig. 6. Significant positive precipitation responses are simulated over the southern TP and in most regions around the dry band mentioned above (such as over northwest China, central Asia and India), which are mainly due to ascent around the area of TP

cooling (Fig. 6d). In Figs. 7c and 7d, a water vapor convergence and vertical ascent band is seen across southern China, the South China Sea, and the North Pacific toward coastal western North America, which matches the band of significant heavy precipitation in Fig. 7a. Equatorward of this moist zone, there are large dry areas over the subtropical North Pacific, associated with water vapor divergence and vertical descent.

Figures 6 and 7 also clearly show that significant seasonal mean responses are simulated over most tropical regions and parts of the North Atlantic, indicating that the TP snow forcing could generate a significant near-global climate impact. Significant SLP and Z500 increases (Figs. 6b,d) are induced over much of the subtropical North Pacific and tropical Indian Ocean, consistent with observed results in Fig. 4b. Northeast trade winds are significantly enhanced over most of the tropical central and western Pacific and Indian oceans (Fig. 6g). In Fig. 7a, significant increased precipitation over the western equatorial tropical Pacific is simulated over the southern part of the climatological intertropical convergence zone (ITCZ, shown by thin black contours), indicating that the ITCZ is shifted southward in that region. In Fig. 7d, the areas of significant 300-hPa vertical ascent or descent over the tropical or subtropical Pacific, respectively, are approximately collocated with the climatological vertical motions shown by thin black contours (dashed lines and blue colors are ascending air).

Figure 8b shows the climatological pressure/latitude cross section of the vectors of meridional horizontal and vertical velocity, and shaded contours show the climatological zonal-mean meridional streamfunction averaged over 100° – 130°E (upward vectors in the troposphere from about 25° – 30°N are part of the midlatitude circulation, and are north of the downward branch of the Hadley circulation, which is weak in this longitude range). Figure 8a shows the pos minus neg responses of the velocity vectors and streamfunction, and indicates that the Hadley circulation between the equator and 30°N is significantly strengthened and shifted northward. The heavy TP snow forcing also simulates significant vertical ascent (Fig. 7d) and positive precipitation (Fig. 7a) and soil moisture anomalies (Fig. 7b) mostly over southern and coastal eastern Africa. Over the western North Atlantic, this forcing induces weak but significant SLP and Z500 reductions (Figs. 6b,d), and a strengthened westerly surface wind and subtropical jet stream around

←

± 4 m for Z500 and Z70. Shading denotes values that are significant at the 95% confidence level from a two-sided Student's t test. Gray-shaded areas in (g) and (h) have significant responses of zonal and/or meridional wind speed at the 95% confidence level. In (h), the NH polar and subtropical jets are shown by overplotting 300-hPa climatological zonal wind speed $\geq 15 \text{ m s}^{-1}$ (red; interval 10 m s^{-1}) from the control run.

FIG. 7. Ensemble-mean MAM averages of the “pos minus neg” responses of four variables: (a) precipitation anomalies, (b) soil moisture anomalies, (c) vertically integrated water vapor flux vectors ($\mathbf{Q}_u$ and $\mathbf{Q}_v$; reference vector is $15 \text{ kg m}^{-1} \text{ s}^{-1}$), and (d) 300-hPa vertical velocity (ascending air has a negative vertical velocity in pressure). Dotted areas in (a), (b), and (d) and shaded areas in (c) denote values that are significant at the 95% confidence level from the two-sided Student’s t test. Black contours in (a) and (d) show the climatological mean precipitation and 300-hPa vertical velocity from the control run, with contour intervals of $100 \text{ mm month}^{-1}$ for precipitation and 0.1 Pa s^{-1} for vertical velocity.

30°N (Figs. 6g,h), which are consistent with observational responses over North Atlantic in Fig. 4b.

Overall, the MCA and model results are consistent over East Asia, the Pacific–North America sector and the tropical Indian and Pacific Oceans. However, observational and model results may still differ for several reasons. First, since MCA only identifies the patterns of two spatial fields that produce the maximum covariance, the MCA patterns in Fig. 4b may not be purely indicative of atmospheric responses to TP snow anomalies. Second, the ensemble averaging process filters out most natural internal variability simulated by individual runs. This is shown in Fig. 4b and other figures by larger observed fluctuations than in the ensemble mean model simulations. Third, atmospheric global circulation model (AGCM) simulations do not allow the atmosphere and

ocean to act as an interactive, coupled system, which may affect some aspects of the model atmosphere’s response to TP snow forcing.

c. Mechanisms of the remote NPO/WP atmospheric response

The local SAT and SLP responses to snow forcing over the TP in Fig. 6 are initialized and maintained by changes in heat and radiative fluxes (Fig. 5). Significant cold SAT and tropospheric atmosphere over the TP are simulated due to increased albedo. Such cooling can produce convergence and descending motion above the plateau (Wu et al. 2014), but ascent around most areas of TP cooling, as shown in Fig. 7d. Since the TP is situated in the subtropics, a tropospheric cold temperature and low height anomaly can be carried or advected to the North Pacific

FIG. 8. (a) Ensemble-mean MAM averages of the zonal mean “pos minus neg” responses (averaged from 100° to 130°E longitude) of vectors composed of meridional velocity [horizontal velocity (m s^{-1}) and vertical velocity (Pa s^{-1})]. Shaded contours show the zonal-mean meridional streamfunction (kg s^{-1}) (b) The corresponding climatological mean cross section of the same variables. Note that vector length scales and contour intervals differ between the panels. Positive values (red shading) indicate clockwise circulation; negative values (blue shading) indicate counterclockwise circulation. Black vectors in (a) indicate significant responses of meridional and/or vertical velocity at the 95% confidence level.

by the prevailing westerlies (Figs. 6c,d,h). A cold TP and North Pacific troposphere tends to enhance the local meridional temperature gradient toward lower latitudes (Fig. 6c) and then (Fig. 6h) strengthens the East Asian subtropical jet. Conversely, a cold TP and North Pacific troposphere tends to weaken the local meridional temperature gradient toward higher latitudes (Fig. 6c) and then (Fig. 6h) weakens the East Asian polar front jet.

In addition to the above thermodynamic processes, dynamical processes are also important for the remote large-

scale NPO/WP response to TP snow forcing. Vectors and contours in Fig. 9a illustrate the seasonal mean response of horizontal and vertical components of 300-hPa stationary Plumb wave activity flux (WAF) (Plumb 1985), respectively. A 10-day low-pass filter is applied to 300-hPa height (Z300) before the horizontal geostrophic velocities are calculated. In Fig. 9a, wave propagation is clearly shown toward the TP snow increase areas from northern China and Mongolia and Southeast Asia, and westward wave propagation is shown from the TP toward the Middle East

(a) 300 hPa Horizontal WAF & Vertical WAF

(b) Z500

(c) 300 hPa EKE

(d) 300 hPa E-P Vectors

(e) Z300 Tendencies

FIG. 9. Ensemble-mean MAM averages of the “pos minus neg” responses of five variables: (a) fields of stationary horizontal wave activity flux (vectors; reference vector is $1.5 \text{ m}^2 \text{ s}^{-2}$) and vertical wave activity flux (contour interval of $0.004 \text{ m}^2 \text{ s}^{-2}$) at 300 hPa, (b) Z500 anomalies (contour interval 4 m), (c) EKE ($\text{m}^2 \text{ s}^{-2}$; contour interval of $1 \text{ m}^2 \text{ s}^{-2}$), (d) horizontal extended transient E-P vectors (reference vector is $2 \text{ m}^2 \text{ s}^{-2}$), and (e) 300-hPa height tendencies (m s^{-1} ; contour interval of $0.5 \times 10^{-5} \text{ m s}^{-1}$). Shaded areas denote values that are significant at the 95% confidence level from a two-sided Student’s t test. Negative contours are dashed, and the zero line is omitted in (a), (b), (c), and (e).

and Europe, indicating that anomalously high TP snow cover weakens horizontal stationary WAF both upstream and downstream of the TP region. Meanwhile, wave propagation is also shown westward toward East Asia and eastward toward North America from an origin in the central North Pacific. In accordance with the WAF definition, the source of the significant WAF in the region of the North Pacific midlatitude jet and storm track, and the associated Z300 anomalies in Fig. 9a, appears to be interactions of transient eddies with the Pacific polar jet. Vertical WAF responses in Fig. 9a indicate that a downward wave activity anomaly over the TP and central North Pacific and an upward wave activity anomaly over northern China, the Japan Sea, and the eastern North Pacific significantly respond to heavy TP snow forcing.

Eddy kinetic energy (EKE) is defined as mean deviations of the 300-hPa horizontal wind speed $(\overline{u'^2 + v'^2})/2$ associated with 2–8-day filtered fluctuations (synoptic-scale weather systems), where u and v are zonal and

meridional wind components, respectively, the overbar is the time average, and primes denote deviations from the corresponding time mean quantities. EKE basically indicates storm-track changes. Figure 9c shows that EKE is significantly reduced over the high-latitude North Pacific into North America, and increased over East China and the midlatitude North Pacific, which indicates that positive TP snow anomalies shift the North Pacific storm track south. Figure 9d shows extended horizontal Eliassen–Palm (E–P) vectors, defined as $[(\overline{v'^2} - \overline{u'^2})/2, -\overline{u'v'}]$ (Trenberth 1986). The transient eddies in the E–P vectors act to accelerate the spring mean flow where the arrows in Fig. 9d are divergent, such as in the enhanced-storm-track region equatorward of approximately 40°N in the North Pacific. Conversely, the transient eddies in the E–P vectors act to decrease the spring mean flow where the arrows in Fig. 9d are convergent, such as in the weakened-storm-track region poleward of approximately 40°N in the North Pacific.

These dynamical effects are consistent with zonal wind changes in Fig. 6h.

Quantitative impacts of the transient eddy forcings on the mean flow at the 300-hPa level are calculated by the geopotential height tendency method below, as outlined in Lau (1988). Where π is defined as convergence of the vorticity flux of transient eddies, the geopotential height tendency $\partial z/\partial t$ is proportional to the inverse Laplacian of π :

$$\frac{\partial z}{\partial t} = \frac{f}{g} \nabla^{-2} \pi, \quad \text{with} \quad (1)$$

$$\pi = \frac{1}{a^2 \cos \theta} \left(\frac{\partial}{\partial \theta} \frac{1}{\cos \theta} \frac{\partial}{\partial \theta} \cos^2 \theta - \frac{1}{\cos \theta} \frac{\partial^2}{\partial \lambda^2} \right) \overline{u'v'} + \frac{1}{a^2 \cos^2 \theta} \frac{\partial^2}{\partial \lambda \partial \theta} \cos \theta (\overline{u'^2} - \overline{v'^2}), \quad (2)$$

where a is Earth's radius, f is the Coriolis parameter, g is the gravitational acceleration, λ is the latitude, and θ is the longitude. Figure 9e shows that transient eddies in MAM due to positive TP snow forcing induce a strong increasing height tendency at 300 hPa over all of the Arctic region, and especially in the area of the Aleutian low and the Russian Far East. Z300 tendencies also show a slight increase over the western subtropical Pacific, and a strong decrease in a band from Mongolia to the central North Pacific. These Z300 tendencies coincide very well with the decreasing Z300 (Fig. 9b) and Z500 (Fig. 6d) across East Asia and all of the North Pacific.

In summary, the growth and maintenance of the negative NPO/WP-like circulation response in MAM to TP positive snow anomalies involves feedback by synoptic transient eddies propagating along the midlatitude jet stream across the whole Eastern Hemisphere, as well as a hemispheric-scale dipole of meridional stationary WAF anomalies centered in the Northern Pacific. This is consistent with previous studies that find a strong dynamical connection between the Pacific zonal flow and the synoptic eddies (Lau 1988; Ren et al. 2008, 2010).

4. Conclusions

This study focuses on seasonal-scale widespread TP snow anomalies that generally originate in the autumn or winter and persist through spring or longer, and finds consistent near-global responses in observations and simulations. Heavy TP snow anomalies first cause significant regional surface cooling and increased precipitation over and surrounding the TP. An induced remote NPO/WP-like teleconnection response in the NH troposphere and stratosphere then extends across the Pacific basin. Cayan (1992) found the surface heat flux

associated with the NPO circulation to have a tripole structure, with consequences for SST. The enhanced Pacific trades in the negative NPO phase (weakened Aleutian low) are linked with tropical/subtropical ocean–atmosphere variability, specifically the Pacific meridional mode (Chiang and Vimont 2004) and ENSO (Vimont et al. 2003). Due mainly to downstream temperature advection, SAT and T500 responses in Fig. 6 show a three-band anomaly structure in the Pacific similar to the pattern of SST anomalies linked with the negative NPO (Cayan 1992; Linkin and Nigam 2008). Other responses extend equatorward into the tropical Pacific and Indian Oceans, poleward to the western Arctic, and westward to the North Atlantic. These results indicate the important role of TP snow anomalies in transient variability in spring when ENSO typically starts to decay and has weak global impacts, and highlight the need for dynamical prediction systems to correctly represent and forecast the evolution of interannual TP snow variability for improving their skill at predicting the surface spring climate in Asia and North America. Our results here also underscore the potential roles of springtime TP snow anomalies in simultaneous ocean–atmosphere interaction in the tropical Pacific and Indian Oceans.

Heavy winter TP snow can lead to a remote hemispheric PNA-like response caused by the eastward propagation of Rossby waves and transient eddy feedbacks (Wu et al. 2011; Liu et al. 2017). Wang et al. (2017) report that a negative WP-like response in summer to increasing albedo in the western TP from late autumn to early spring of the next year is simulated (their Fig. 10e). Therefore, TP snow anomalies in nearly all seasons are important sources of variability of the large-scale atmospheric circulation at seasonal and interannual time scales, and on ocean–atmosphere variability in the tropical/subtropical Pacific. On a multidecadal scale, TP snow was generally light during the 1960s and 1970s, heavy in the 1980s and 1990s, and light since 2000 (Zhang et al. 2004; Si and Ding 2013). Such decadal TP snow variations may play an important role in decadal teleconnections including the Pacific decadal oscillation and Arctic Oscillation, and further research is needed about possible effects of long-term TP snow trends on future global climate changes.

At present, passive microwave remote sensing is the only efficient way to monitor temporal and spatial variations in snow depth at a large scale. Satellite mapping of snow depth and SWE has lower accuracy than for SCE, especially in mountainous and heavily forested areas (Vaughan et al. 2013). Previous research finds that existing snow SCE and SWE products have uncertainties over the TP (Dai et al. 2017). Although CLM4.0

combines realistic radiative, ecological and hydrologic processes (Oleson et al. 2010), snow processes are a major weakness in land surface models (Dirmeyer et al. 2006). Snow parameterization in such complex terrain as the TP is always challenging. Since TP snow forcing is prescribed from satellite SCE and SWE, CLM4.0 ice-snow albedo calculations can lead to uncertainties in surface diabatic heating. However, modeled hemispheric responses are induced by physically based large-scale dynamics. Observational and modeled results are remarkably consistent, which should not be simply a random coincidence. Our previous simulations with the same models and experiment design reproduce the link of large positive autumn and winter TP SCE anomalies to the PNA teleconnection in winter (Liu et al. 2017). Therefore, we conclude that the main conclusions here are not distorted by uncertainties of satellite SCE and SWE over the TP, or by CLM4.0 snow-albedo calculations. However, the quality, appropriateness, and reliability of observational data is still an important caveat for modeling studies, because data limitations can affect the realism of both input scenarios and model outputs. This suggests the importance of continued development of long-term observational satellite TP snow datasets.

Acknowledgments. This work is funded by the National Key Scientific Research Plan of China (Grants 91837206 and 91937302) and the National Natural Science Foundation of China (Grants 41375076) and is also supported by the Jiangsu Collaborative Innovation Center for Climate Change and the CMA-NJU Joint Laboratory for Climate Prediction Studies. All simulations were carried out at National Supercomputer Center in Tianjin, and the calculations were performed on TianHe-1 (A).

REFERENCES

- Alexander, M., R. Tomas, C. Deser, and D. M. Lawrence, 2010: The atmospheric response to projected terrestrial snow changes in the late twenty-first century. *J. Climate*, **23**, 6430–6437, <https://doi.org/10.1175/2010JCLI3899.1>.
- Allen, R. J., and C. S. Zender, 2010: Effects of continental-scale snow albedo anomalies on the wintertime Arctic oscillation. *J. Geophys. Res.*, **115**, D23105, <https://doi.org/10.1029/2010JD014490>.
- Anderson, E. A., 1976: A point energy and mass balance model of a snow cover. National Weather Service Office of Hydrology NOAA Tech. Rep. NWS 19, 150 pp., <https://repository.library.noaa.gov/view/noaa/6392>.
- Armstrong, R., M. J. Brodzik, K. Knowles, and M. Savoie, 2005: Global Monthly EASE-Grid Snow Water Equivalent Climatology, version 1. National Snow and Ice Data Center Distributed Active Archive Center, accessed August 2016, <https://doi.org/10.5067/KJVERY3MIBPS>.
- Bamzai, A. S., and J. Shukla, 1999: Relation between Eurasian snow cover, snow depth, and the Indian summer monsoon: An observational study. *J. Climate*, **12**, 3117–3132, [https://doi.org/10.1175/1520-0442\(1999\)012<3117:RBESCS>2.0.CO;2](https://doi.org/10.1175/1520-0442(1999)012<3117:RBESCS>2.0.CO;2).
- Blanford, H. F., 1884: On the connexion of the Himalaya snowfall with dry winds and seasons of drought in India. *Proc. Roy. Soc. London*, **37**, 3–22, <https://doi.org/10.1098/rspl.1884.0003>.
- Brodzik, M., and R. Armstrong, 2013: Northern Hemisphere EASE-Grid 2.0 Weekly Snow Cover and Sea Ice Extent, version 4. NASA National Snow and Ice Data Center, accessed August 2016, <https://doi.org/10.5067/P7O0HGJLYUQU>.
- Cayan, D. R., 1992: Latent and sensible heat flux anomalies over the northern oceans: The connection to monthly atmospheric circulation. *J. Climate*, **5**, 354–369, [https://doi.org/10.1175/1520-0442\(1992\)005<0354:LASHFA>2.0.CO;2](https://doi.org/10.1175/1520-0442(1992)005<0354:LASHFA>2.0.CO;2).
- Chiang, J. C., and D. J. Vimont, 2004: Analogous Pacific and Atlantic meridional modes of tropical atmosphere–ocean variability. *J. Climate*, **17**, 4143–4158, <https://doi.org/10.1175/JCLI4953.1>.
- Clark, M. P., and M. C. Serreze, 2000: Effects of variations in East Asian snow cover on modulating atmospheric circulation over the North Pacific Ocean. *J. Climate*, **13**, 3700–3710, [https://doi.org/10.1175/1520-0442\(2000\)013<3700:EOVIEA>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<3700:EOVIEA>2.0.CO;2).
- Cohen, J., and D. Rind, 1991: The effect of snow cover on the climate. *J. Climate*, **4**, 689–706, [https://doi.org/10.1175/1520-0442\(1991\)004<0689:TEOSCO>2.0.CO;2](https://doi.org/10.1175/1520-0442(1991)004<0689:TEOSCO>2.0.CO;2).
- , and D. Entekhabi, 1999: Eurasian snow cover variability and Northern Hemisphere climate predictability. *Geophys. Res. Lett.*, **26**, 345–348, <https://doi.org/10.1029/1998GL900321>.
- , and C. Fletcher, 2007: Improved skill for Northern Hemisphere winter surface temperature predictions based on land-atmosphere fall anomalies. *J. Climate*, **20**, 4118–4132, <https://doi.org/10.1175/JCLI4241.1>.
- Czaja, A., and C. Frankignoul, 2002: Observed impact of Atlantic SST anomalies on the North Atlantic Oscillation. *J. Climate*, **15**, 606–623, [https://doi.org/10.1175/1520-0442\(2002\)015<0606:OIOASA>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<0606:OIOASA>2.0.CO;2).
- Dai, L., T. Che, Y. Ding, and X. Hao, 2017: Evaluation of snow cover and snow depth on the Qinghai-Tibetan Plateau derived from passive microwave remote sensing. *Cryosphere*, **11**, 1933–1948, <https://doi.org/10.5194/tc-11-1933-2017>.
- Dai, Y., and Q. Zeng, 1997: A land surface model (IAP94) for climate studies. Part I: formulation and validation in off-line experiments. *Adv. Atmos. Sci.*, **14**, 433–460, <https://doi.org/10.1007/s00376-997-0063-4>.
- Dirmeyer, P. A., K. D. Randal, and Z. Guo, 2006: Do global models properly represent the feedback between land and atmosphere? *J. Hydrol.*, **7**, 1177–1198, <https://doi.org/10.1175/JHM532.1>.
- Fasullo, J., 2004: A stratified diagnosis of the Indian monsoon–Eurasian snow cover relationship. *J. Climate*, **17**, 1110–1122, [https://doi.org/10.1175/1520-0442\(2004\)017<1110:ASDOTI>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<1110:ASDOTI>2.0.CO;2).
- Fletcher, C. G., S. C. Hardiman, and P. J. Kushner, 2009: The dynamical response to snow cover perturbations in a large ensemble of atmospheric GCM integrations. *J. Climate*, **22**, 1208–1222, <https://doi.org/10.1175/2008JCLI2505.1>.
- Gong, G., D. Entekhabi, and J. Cohen, 2003: Modeled Northern Hemisphere winter climate response to realistic Siberian snow anomalies. *J. Climate*, **16**, 3917–3931, [https://doi.org/10.1175/1520-0442\(2003\)016<3917:MNHWCRC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<3917:MNHWCRC>2.0.CO;2).
- Groisman, P. Y., T. Karl, R. W. Knight, and G. L. Stenchikov, 1994: Changes of snow cover, temperature and radiative heat balance over the Northern Hemisphere. *J. Climate*, **7**, 1633–1656, [https://doi.org/10.1175/1520-0442\(1994\)007<1633:COSCTA>2.0.CO;2](https://doi.org/10.1175/1520-0442(1994)007<1633:COSCTA>2.0.CO;2).

- Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471, [https://doi.org/10.1175/1520-0477\(1996\)077<0437:TNYRP>2.0.CO;2](https://doi.org/10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2).
- Lau, N.-C., 1988: Variability in the observed midlatitude storm tracks in relation to low-frequency changes in the circulation pattern. *J. Atmos. Sci.*, **45**, 2718–2743, [https://doi.org/10.1175/1520-0469\(1988\)045<2718:VOTOMS>2.0.CO;2](https://doi.org/10.1175/1520-0469(1988)045<2718:VOTOMS>2.0.CO;2).
- Lin, H., and Z. Wu, 2011: Contribution of the autumn Tibetan Plateau snow cover to seasonal prediction of North American winter temperature. *J. Climate*, **24**, 2801–2813, <https://doi.org/10.1175/2010JCLI3889.1>.
- Lin, P., J. Wei, Z.-L. Yang, Y. Zhang, and K. Zhang, 2016: Snow data assimilation-constrained land initialization improves seasonal temperature prediction. *Geophys. Res. Lett.*, **43**, 11 423–11 432, <https://doi.org/10.1002/2016GL070966>.
- Linkin, M. E., and S. Nigam, 2008: The North Pacific Oscillation–West Pacific teleconnection pattern: Mature-phase structure and winter impacts. *J. Climate*, **21**, 1979–1997, <https://doi.org/10.1175/2007JCLI2048.1>.
- Liu, S., Q. Wu, X. Ren, Y. Yao, S. R. Schroeder, and H. Hu, 2017: Modeled Northern Hemisphere autumn and winter climate responses to realistic Tibetan Plateau and Mongolia snow anomalies. *J. Climate*, **30**, 9435–9454, <https://doi.org/10.1175/JCLI-D-17-0117.1>.
- Mote, T. L., and E. R. Kutney, 2012: Regions of autumn Eurasian snow cover and associations with North American winter temperatures. *Int. J. Climatol.*, **32**, 1164–1177, <https://doi.org/10.1002/joc.2341>.
- Namias, J., 1985: Some empirical evidence for the influence of snow cover on temperature and precipitation. *Mon. Wea. Rev.*, **113**, 1542–1553, [https://doi.org/10.1175/1520-0493\(1985\)113<1542:SEEFIT>2.0.CO;2](https://doi.org/10.1175/1520-0493(1985)113<1542:SEEFIT>2.0.CO;2).
- Neale, R., J. Richter, S. Park, P. H. Lauritzen, S. J. Vavrus, P. J. Rasch, and M. Zhang, 2013: The mean climate of the Community Atmosphere Model (CAM4) in forced SST and fully coupled experiments. *J. Climate*, **26**, 5150–5168, <https://doi.org/10.1175/JCLI-D-12-00236.1>.
- Oleson, K. W., and Coauthors, 2010: Technical description of version 4.5 of the Community Land Model (CLM). NCAR Tech. Note NCAR/TN-478+STR, 257 pp., <https://doi.org/10.5065/D6FB50WZ>.
- Orsolini, Y. J., and N. G. Kvamstø, 2009: Role of Eurasian snow cover in wintertime circulation: Decadal simulations forced with satellite observations. *J. Geophys. Res.*, **114**, D19108, <https://doi.org/10.1029/2009JD012253>.
- Plumb, R., 1985: On the three-dimensional propagation of stationary waves. *J. Atmos. Sci.*, **42**, 217–229, [https://doi.org/10.1175/1520-0469\(1985\)042<0217:OTTDPO>2.0.CO;2](https://doi.org/10.1175/1520-0469(1985)042<0217:OTTDPO>2.0.CO;2).
- Pu, Z., L. Xu, and V. V. Salomonson, 2007: MODIS/Terra observed seasonal variations of snow cover over the Tibetan Plateau. *Geophys. Res. Lett.*, **34**, L06706, <https://doi.org/10.1029/2007GL029262>.
- Ren, X., Y. Zhang, and Y. Xiang, 2008: Connections between wintertime jet stream variability, oceanic surface heating, and transient eddy activity in the North Pacific. *J. Geophys. Res.*, **113**, D21119, <https://doi.org/10.1029/2007JD009464>.
- , X. Yang, and C. Chu, 2010: Seasonal variations of the synoptic-scale transient eddy activity and polar front jet over East Asia. *J. Climate*, **23**, 3222–3233, <https://doi.org/10.1175/2009JCLI3225.1>.
- Senan, R., and Coauthors, 2016: Impact of springtime Himalayan–Tibetan Plateau snowpack on the onset of the Indian summer monsoon in coupled seasonal forecasts. *Climate Dyn.*, **47**, 2709–2725, <https://doi.org/10.1007/s00382-016-2993-y>.
- Shukla, J., and D. A. Mooley, 1987: Empirical prediction of the summer monsoon rainfall over India. *Mon. Wea. Rev.*, **115**, 695–704, [https://doi.org/10.1175/1520-0493\(1987\)115<0695:EPOTSM>2.0.CO;2](https://doi.org/10.1175/1520-0493(1987)115<0695:EPOTSM>2.0.CO;2).
- Si, D., and Y. Ding, 2013: Decadal change in the correlation pattern between the Tibetan Plateau winter snow and the East Asian summer precipitation during 1979–2011. *J. Climate*, **26**, 7622–7634, <https://doi.org/10.1175/JCLI-D-12-00587.1>.
- Smith, K. L., P. J. Kushner, and J. Cohen, 2011: The role of linear interference in northern annular mode variability associated with Eurasian snow cover extent. *J. Climate*, **24**, 6185–6202, <https://doi.org/10.1175/JCLI-D-11-00055.1>.
- Thompson, D. W. J., and J. M. Wallace, 1998: The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.*, **25**, 1297–1300, <https://doi.org/10.1029/98GL00950>.
- Trenberth, K. E., 1986: An assessment of the impact of transient eddies on the zonal flow during a blocking episode using localized Eliassen–Palm flux diagnostics. *J. Atmos. Sci.*, **43**, 2070–2087, [https://doi.org/10.1175/1520-0469\(1986\)043<2070:AAOTIO>2.0.CO;2](https://doi.org/10.1175/1520-0469(1986)043<2070:AAOTIO>2.0.CO;2).
- Turner, A. G., and J. M. Slingo, 2011: Using idealized snow forcing to test teleconnections with the Indian summer monsoon in the Hadley Centre GCM. *Climate Dyn.*, **36**, 1717–1735, <https://doi.org/10.1007/s00382-010-0805-3>.
- Vaughan, D. G., and Coauthors, 2013: Observations: Cryosphere. *Climate Change 2013: The Physical Science Basis*, T. F. Stocker et al., Eds., Cambridge University Press, 317–382.
- Vimont, D. J., J. M. Wallace, and D. S. Battisti, 2003: The seasonal footprinting mechanism in the Pacific: Implications for ENSO. *J. Climate*, **16**, 2668–2675, [https://doi.org/10.1175/1520-0442\(2003\)016<2668:TSFMIT>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<2668:TSFMIT>2.0.CO;2).
- Wallace, J. M., and D. S. Gutzler, 1981: Teleconnections in the geopotential height field during the Northern Hemisphere winter. *Mon. Wea. Rev.*, **109**, 784–812, [https://doi.org/10.1175/1520-0493\(1981\)109<0784:TITGHF>2.0.CO;2](https://doi.org/10.1175/1520-0493(1981)109<0784:TITGHF>2.0.CO;2).
- Walland, D. J., and I. Simmonds, 1996: Modeled atmospheric response to changes in Northern Hemisphere snow cover. *Climate Dyn.*, **13**, 25–34, <https://doi.org/10.1007/s003820050150>.
- Walsh, J. E., and B. Ross, 1988: Sensitivity of 30-day dynamical forecasts to continental snow cover. *J. Climate*, **1**, 739–754, [https://doi.org/10.1175/1520-0442\(1988\)001<0739:SODDFT>2.0.CO;2](https://doi.org/10.1175/1520-0442(1988)001<0739:SODDFT>2.0.CO;2).
- Wang, B., Q. Bao, B. Hoskins, G. Wu, and Y. Liu, 2008: Tibetan Plateau warming and precipitation changes in East Asia. *Geophys. Res. Lett.*, **35**, L14702, <https://doi.org/10.1029/2008GL034330>.
- Wang, C. H., K. Yang, Y. Li, D. Wu, and Y. Bo, 2017: Impacts of spatiotemporal anomalies of Tibetan Plateau snow cover on summer precipitation in eastern China. *J. Climate*, **30**, 885–903, <https://doi.org/10.1175/JCLI-D-16-0041.1>.
- Wang, Z., R. Wu, S. Chen, G. Huang, G. Liu, and L. Zhu, 2018: Influence of western Tibetan Plateau summer snow cover on East Asian summer rainfall. *J. Geophys. Res. Atmos.*, **123**, 2371–2386, <https://doi.org/10.1002/2017JD028016>.
- Webster, P. J., V. O. Magana, T. N. Palmer, J. Shukla, R. A. Tomas, M. Yanai, and T. Yasunari, 1998: Monsoons: Processes, predictability, and the prospects for prediction. *J. Geophys. Res.*, **103**, 14 451–14 510, <https://doi.org/10.1029/97JC02719>.
- Wu, G., and Coauthors, 2014: Tibetan Plateau climate dynamics: Recent research progress and outlook. *Natl. Sci. Rev.*, **2**, 100–116, <https://doi.org/10.1093/nsr/nwu045>.
- Wu, Q., H. Hu, and L. Zhang, 2011: Observed influences of autumn–early winter Eurasian snow cover anomalies on the

- hemispheric PNA-like variability in winter. *J. Climate*, **24**, 2017–2023, <https://doi.org/10.1175/2011JCLI4236.1>.
- Wu, T. W., and Z. A. Qian, 2003: The relation between the Tibetan winter snow and the Asian summer monsoon and rainfall: An observational investigation. *J. Climate*, **16**, 2038–2051, [https://doi.org/10.1175/1520-0442\(2003\)016<2038:TRBTTW>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<2038:TRBTTW>2.0.CO;2).
- Xu, X., J. Guo, T. Koike, Y. Liu, X. Shi, F. Zhu, and S. Zhang, 2012: “Downstream effect” of winter snow cover over the eastern Tibetan Plateau on climate anomalies in East Asia. *J. Meteor. Soc. Japan*, **90**, 113–130, <https://doi.org/10.2151/jmsj.2012-C08>.
- Xiao, Z., and A. Duan, 2016: Impacts of Tibetan Plateau snow cover on the interannual variability of the East Asian summer monsoon. *J. Climate*, **29**, 8495–8514, <https://doi.org/10.1175/JCLI-D-16-0029.1>.
- Yanai, M., C. Li, and Z. Song, 1992: Seasonal heating of the Tibetan Plateau and its effects on the evolution of the Asian summer monsoon. *J. Meteor. Soc. Japan*, **70**, 319–351, https://doi.org/10.2151/jmsj1965.70.1B_319.
- Yasunari, T., A. Kitoh, and T. Tokioka, 1991: Local and remote responses to excessive snow mass over Eurasia appearing in the northern spring and summer climate—A study with the MRI GCM. *J. Meteor. Soc. Japan*, **69**, 473–487, https://doi.org/10.2151/jmsj1965.69.4_473.
- Zhang, Y., T. Li, and B. Wang, 2004: Decadal change of the spring snow depth over the Tibetan Plateau: The associated circulation and influence on the East Asian summer monsoon. *J. Climate*, **17**, 2780–2793, [https://doi.org/10.1175/1520-0442\(2004\)017<2780:DCOTSS>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<2780:DCOTSS>2.0.CO;2).
- Zhao, H., and G. W. K. Moore, 2004: On the relationship between Tibetan snow cover, the Tibetan plateau monsoon and the Indian summer monsoon. *Geophys. Res. Lett.*, **31**, L14204, <https://doi.org/10.1029/2004GL020040>.