The Summer Snow Cover Anomaly over the Tibetan Plateau and Its Association with Simultaneous Precipitation over the Mei-yu–Baiu region

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(Received 9 September 2013; revised 09 December 2013; accepted 27 December 2013)

ABSTRACT

The summer snow anomalies over the Tibetan Plateau (TP) and their effects on climate variability are often overlooked, possibly due to the fact that some datasets cannot properly capture summer snow cover over high terrain. The satellite-derived Equal-Area Scalable Earth grid (EASE-grid) dataset shows that snow still exists in summer in the western part and along the southern flank of the TP. Analysis demonstrates that the summer snow cover area proportion (SCAP) over the TP has a significant positive correlation with simultaneous precipitation over the mei-yu–baiu (MB) region on the interannual time scale. The close relationship between the summer SCAP and summer precipitation over the MB region could not be simply considered as a simultaneous response to the Silk Road pattern and the SST anomalies in the tropical Indian Ocean and tropical central-eastern Pacific. The SCAP anomaly has an independent effect and may directly modulate the land surface heating and, consequently, vertical motion over the western TP, and concurrently induce anomalous vertical motion over the North Indian Ocean via a meridional vertical circulation. Through a zonal vertical circulation over the tropics and a Kelvin wave-type response, anomalous vertical motion over the North Indian Ocean may result in an anomalous high over the western North Pacific and modulate the convective activity in the western Pacific warm pool, which stimulates the East Asia–Pacific (EAP) pattern and eventually affects summer precipitation over the MB region.

Key words: snow cover, Tibetan Plateau, mei-yu, baiu, precipitation

Citation: Liu, G., R. G. Wu, Y. Z. Zhang, and S. L. Nan, 2014: The summer snow cover anomaly over the Tibetan Plateau and its association with simultaneous precipitation over the mei-yu–baiu region. *Adv. Atmos. Sci.*, **31**(4), 755–764, doi: 10.1007/s00376-013-3183-z.

1. Introduction

A quasi-stationary rainband extending from eastern China to Japan, called “mei-yu” in China and “baiu” in Japan, is an important climate phenomenon of East Asia in early summer (Tao and Chen, 1987; Ninomiya and Murakami, 1987). The mei-yu–baiu (hereafter MB) front brings dominant precipitation over East Asia and even determines the distribution of precipitation in the whole summer (June–August, JJA). Anomalous summer precipitation over the MB region may cause drought or flooding and exert important societal and economic impacts on the highly populated East Asian countries. Therefore, the mechanisms and short-term climate forecasts of MB precipitation are of great interest to climate scientists, governments and the public.

Precipitation in the MB region, which is closely associated with the variability of the East Asian summer monsoon (EASM), is affected by anomalous nonlocal lower boundary conditions, such as SST, sea ice, snow, and soil moisture (Chang et al., 2000; Chen and Wu, 2000; Kripalani et al., 2002; Yang and Lau, 2004; Zhao et al., 2004; Ding and Chan, 2005; Zuo and Zhang, 2007; Wang et al., 2012). Of the above-mentioned boundary conditions, snow anomalies affect the thermal characteristics of the land surface and lower troposphere through modulating radiation, moisture, and the energy budget (Namias, 1985; Walsh et al., 1985; Cohen and Rind, 1991; Karl et al., 1993; Groisman et al., 1993). Thus, Tibetan Plateau (TP) heating induced by snow-cover change can trigger large-scale anomalous atmospheric circulation, consequently leading to anomalous Asian summer monsoon and associated monsoon rainfall (Yanai et al., 1992; Webster et al., 1998; Wu and Qian, 2000; Zhang et al., 2004; Wang et al., 2008).

There have been some previous investigations on the relationship between the TP snow anomaly and East Asian monsoon rainfall (including MB rainfall). For example, Chen and Wu (2000) found the winter–spring TP snow depth is posi-

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tively correlated with boreal summer rainfall in central China along the middle and lower reaches of the Yangtze River (namely the mei-yu region of China), and negatively correlated with that in northern and southern China. Zhang et al. (2004) argued that the interdecadal variation of snow depth over the eastern TP in March–April is connected with a wetter (drier) summer climate in the Yangtze River valley and a dryer (wetter) one over the southeast coast of China and the Indochina peninsula. Wu and Kirtman (2007) indicated that spring snow cover over the TP shows a moderate positive correlation with spring rainfall in southern China. Zhao et al. (2007) pointed out that an increase in the number of snow-covered days (SCDs) in spring may result in a weakening of the EASM and a decrease of rainfall over the Yangtze and Huaihe Rivers and an increase in southeastern China.

The above studies mainly focused on the relationship between preceding winter–spring snow over the TP and the following spring or summer rainfall. It is important to note that anomalous summer heating resulting from boundary conditions over the TP obviously plays a more direct role in the variability of EASM rainfall. Zhao and Chen (2001) found a remarkable correlation between the summer TP heat source and simultaneous rainfall over the Yangtze River valley. Wang et al. (2008) indicated that atmospheric heating induced by rising TP temperatures can enhance East Asian subtropical frontal rainfall. The memory of anomalous winter snow in the climate system resides in the wetness of the underlying soil as snow melts during the spring and summer seasons. This notion may generally explain the lingering effect of preceding snow anomalies on summer Indian monsoon rainfall (Shukla and Mooley, 1987; Bamzi and Shukla, 1999). This was also supported by Zhao et al. (2007) who considered that soil moisture anomalies in May–June may act as a bridge linking the previous season's snow over the TP and the subsequent EASM rainfall.

However, it should be noted that although snow cover clearly melts and disappears from winter and spring to summer over the main body of the eastern TP, it may persist through summer at high altitudes, such as in the western and southern TP where there are large mountain ridges (Pu et al., 2007; Wu et al., 2012b). The summer snow cover over the western and southern TP has a potential contribution to the nonhomogeneous pattern of summer atmospheric heating over the TP. For example, summer atmospheric heat sources show an uneven distribution with stronger heating along the southern and southeastern edges and weaker heating over the northern part of the TP (Duan and Wu, 2003). Meanwhile, there are different features in summer atmospheric heating between the eastern TP and the western and southern TP, with a stronger latent heating relative to sensible heating over the eastern TP and a stronger sensible heating relative to latent heating over the western and southern TP (Table 1 in Zhao and Chen, 2001). The summer snow cover over the western and southern TP may to some extent modulate simultaneous atmospheric heating in situ, which may affect EASM rainfall in a way that is different from the atmospheric heating over the eastern main body of the TP. Therefore the relationship between summer snow cover and simultaneous rainfall over East Asia and the related mechanism are worthy of investigation

In summer, the atmospheric heat source along the southern edge of the TP is one of the maximum heating centers (Duan and Wu, 2003). Additionally, numerical simulations have shown that Himalayan heating plays a crucial role in modulating Indian monsoon rainfall (Wu et al., 2012a). These results imply that although it covers a relatively small region over the western and southern TP, summer snow cover may have a more important impact on Asian climate than expected. Recently, the simultaneous effects of summer snow cover over the TP have been found important for the decadal to interdecadal variations of heat waves over northern China and for El Niño–Southern Oscillation (ENSO) teleconnections (Wu et al., 2012b, c). The pronounced effects of summer snow cover over the TP motivated us to further investigate the relationship between summer snow cover and simultaneous rainfall over East Asia.

The organization of the rest of the text is as follows. In section 2, the datasets used in the present study are described. The relationship between summer snow cover over the TP and simultaneous rainfall over East Asia is investigated in section 3, and the associated mechanism is explored in section 4. A summary and discussion are presented in section 5.

2. Datasets

Snow datasets used in the present study include monthly satellite-derived snow water equivalent (SWE) from November 1978 through May 2007 (Armstrong et al., 2005), obtained from the National Snow and Ice Data Center (NSIDC). The data are gridded to the northern and southern 25 km Equal-Area Scalable Earth Grids (EASE-Grids). The SWE in the dataset is derived from the Scanning Multichannel Microwave Radiometer (SMMR) and selected Special Sensor Microwave/Imagers (SSM/I). With respect to those pixels with no microwave-derived SWE, the percentage frequency of visible snow derived from weekly data is added in the dataset as a supplement (more information about this dataset is available online at http://nsidc.org/). For convenience, the data were converted to regular 721×721 grids by setting the value at a regular grid to that at the nearest EASE grid. We also utilize monthly station snow depth data at 174 stations in the TP within the territory of China, provided by the National Meteorological Information Center, China Meteorological Administration. The 20th century reanalysis monthly percentage of snow coverage and snow depth data at 192×94 Gaussian grids (Compo et al., 2006), obtained from the National Oceanic and Atmospheric Administration–Cooperative Institute for Research in Environmental Sciences (NOAA–CIRES) Climate Diagnostics Center, Boulder, Colorado (available online at http://www.cdc.noaa.gov/), are also used for the comparison of snow variation over the TP.

The precipitation dataset used in the present study is the monthly mean Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin, 1997), which is available on $2.5^\circ \times 2.5^\circ$ grids starting from January 1979. The monthly mean atmospheric variables, including geopotential height, winds, vertical velocity etc. were obtained from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al., 1996). The land skin temperature data were obtained from the NCEP–Department of Energy (DOE) reanalysis version 2 (Kanamitsu et al., 2002). The NOAA extended reconstructed version 3b monthly mean SSTs (Smith et al., 2008) are also used in this study.

3. Relationship between the summer snow cover area proportion and precipitation

Since one of the main purposes of the present study is to demonstrate the effect of summer snow anomalies over the TP, we first compare summer snow variables in the different datasets. For the satellite-derived EASE-grid dataset, the following process was carried out to calculate the snow cover extent. The snow cover grids with SWE over 1 mm and snow cover percentage over 35% were assigned the number “1”, and those grids with no snow cover were denoted using the number “0”. In this way, we were able to calculate the snow cover extent over the TP based on synthetic information of SWE and snow cover percentage of visible snow in those pixels with no microwave-derived SWE.

Figure 1 shows the distribution of the climatological percentage of summer and winter snow cover (based on the EASE-grid dataset) over the TP for the period 1979–2006. In winter (Fig. 1b), the area with snow cover years more than 10% of total years occupies almost the entire TP region, and the area with snow cover years more than 50% also accounts for a large proportion of this region. In summer (Fig. 1a), the area with snow cover years more than 10% shrinks in the western region and southern flank of the TP, and the area with over 50% snow cover years is along the Himalayan Mountains. Although the snow area shrinks in summer, it still covers a large region and accounts for approximately 23% of the winter snow area in the TP region. This indicates that the effects of summer snow cover over the TP should not be ignored. However, other datasets show that snow almost completely melts over the TP in summer, with the snow area accounting for approximately 0.7% (5%) of winter snow area based on the 20th century reanalysis snow cover (depth) and 2% based on station observations of snow depth. It appears that the EASE-grid dataset can represent summer snow cover over the TP better than the other datasets mentioned above. Therefore, the former was chosen in the present study.

To quantitatively measure the variation of snow cover over the TP, an index is defined as the snow cover area proportion (SCAP) of the TP region (25°–43°N, 64°–105°E), hereafter called the SCAP index. Note that clear linear decreases can be identified in the summer SCAP index (not shown). Therefore, to investigate the relationship between the summer SCAP index and precipitation on the interannual time scale, we first removed the linear trend in the summer SCAP index and then calculated the correlation of the detrended SCAP index with simultaneous precipitation over East Asia (Fig. 2). Hereafter, the linear trends in all variables have been removed unless otherwise stated. Figure 2a shows that significant negative correlation appears over the region to the east of the Philippines, and positive correlation is distinctly seen over central-eastern China around 30°N (i.e. the mei-yu region) and central Japan (i.e. the baiu region). According to Fig. 2a, summer precipitation averaged over the positive correlation regions, i.e. the mei-yu region (28°–35°N, 107°–120°E) and the baiu region (35°–43°N, 128°–154°E), is considered as an index to represent the variation in summer precipitation over the MB band, hereafter called the MB precipitation index. The MB precipitation index and the SCAP index have a correlation coefficient of 0.62, which is significant at the 99.9% confidence level (Fig. 2b). In fact, summer precipitation over the MB region does not show a clear linear trend during 1979–2006. The original MB precipitation index is still linked to the SCAP index with a correlation coefficient of 0.61. The results demonstrate that summer snow cover over the TP is closely related to the variation in summer precipitation over the MB band, particularly on the interannual time scale. Corresponding to higher (lower) summer snow cover over the TP, summer precipitation increases (decreases) over the MB region.

4. Potential mechanism

Before exploring the potential mechanism of the relationship between the SCAP index and summer precipitation, we first examine the atmospheric circulation anomalies related to summer precipitation over the MB region. The correlation between the summer MB precipitation index and simultaneous 500-hPa geopentential height shows a positive–negative–positive significant correlation pattern from the tropical western Pacific centered on the region around Taiwan and the Philippines, through central Japan, to the high latitudes of East Asia (Fig. 3a). A large anomalous anticyclone appears over the western Pacific in the regressed 850-hPa winds (Fig. 3b), which indicates a stronger western Pacific subtropical high (WPSH). Along the northwestern rim of the stronger WPSH, a stronger anomalous southwesterly flow prevails and transports more moisture to China and Japan and feeds the MB rainband (Akiyama, 1973; Kodama, 1992). Moreover, an anomalous cyclone over central Japan (Fig. 3b), which is linked to the anomalous Bonin (Ogasawara) high (Ninomiya and Akiyama, 1992; Enomoto et al., 2003), may bring more cold air along its western and southwestern rim to converge with warm and wet air along the western and northwestern rim of the WPSH and favor summer precipitation over the MB region. The above anomalous circulations are in good agreement with the East Asia–Pacific (EAP) pattern (Huang and Li, 1988) or the Pacific–Japan (PJ) teleconnection pattern (Nitta, 1987). The EAP/PJ pattern is considered as an important factor for the interannual variation of the EASM and

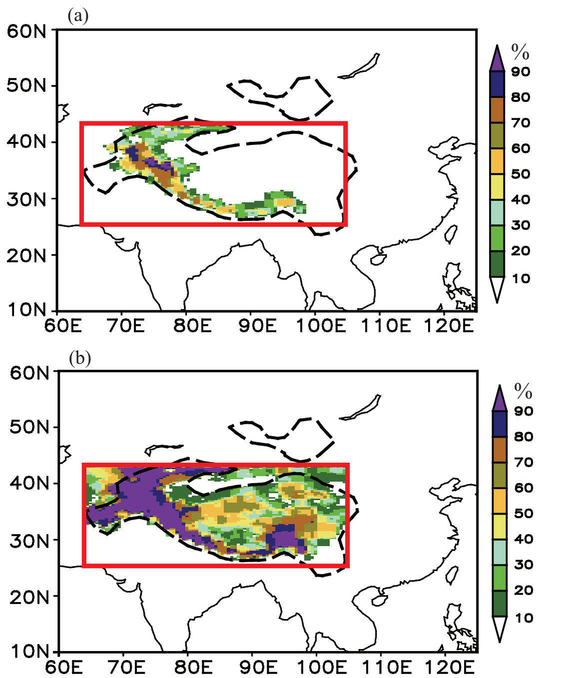

Fig. 1. Climatological percentage (%) distributions of (a) summer and (b) winter snow cover over the TP for the period 1979–2006. The black dashed lines indicate the topographic contour of 1500 m, and the TP region (25°–43°N, 64°–105°E) is indicated by the red boxes.

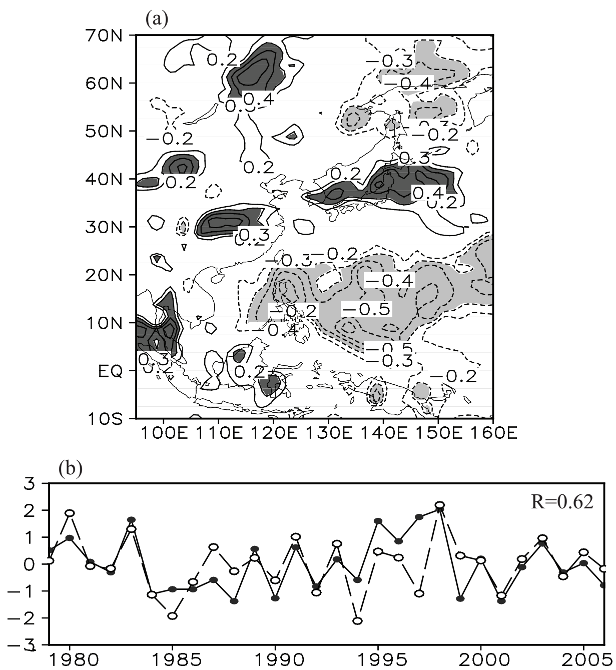

Fig. 2. (a) Spatial distribution of correlation coefficients between the summer SCAP index and summer precipitation for the period 1979–2006, with the shaded areas denoting correlation significant at the 90% confidence level. (b) The normalized time series of the summer SCAP index (solid line with dots) and the simultaneous MB precipitation index (dashed line with circles); their correlation coefficient R is 0.62.

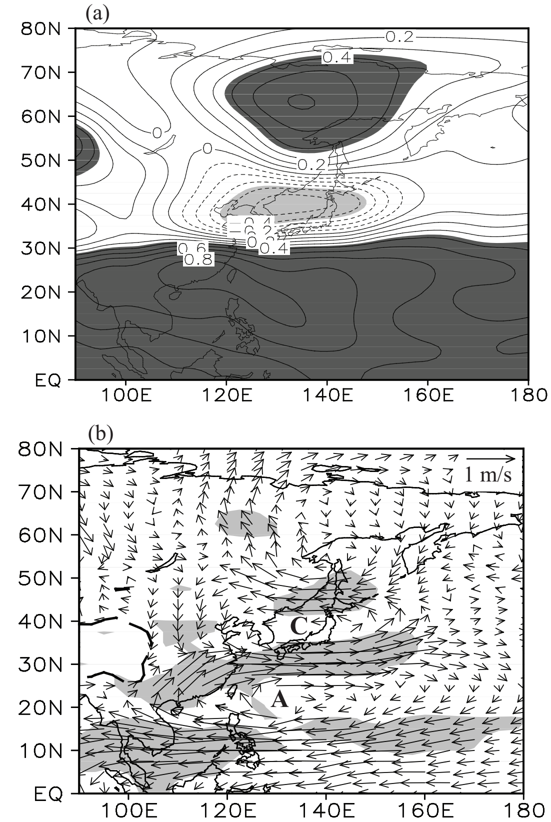

Fig. 3. (a) Correlation of summer 500-hPa geopotential height and (b) regressed summer 850-hPa winds with the summer MB precipitation index. The areas significant at the 95% confidence level are shaded. The "C" and "A" indicate anomalous cyclone and anticyclone, respectively.

associated summer precipitation over the MB region (Nitta, 1987; Huang and Li, 1988; Huang, 2004).

The correlation between the summer SCAP index and simultaneous 500-hPa geopentential height and 850-hPa winds (Fig. 4) shows a pattern very similar to Fig. 3. This implies that the EAP (or PJ) pattern should be an important linkage responsible for the close relationship between summer snow cover anomalies and summer precipitation over the MB region.

The meridional propagation of quasi-stationary planetary Rossby waves triggered by anomalous convective activity in the western Pacific warm pool is an important factor in the formation of the EAP (or PJ) pattern (Huang and Li, 1987; Nitta, 1987; Huang and Sun, 1992). Additionally, the quasi-zonal propagations of Rossby waves over high- and midlatitude regions of the Eurasian continent and over the Asian jet region are responsible for generating basic patterns of high- and midlatitude anomaly centers of the EAP pattern (Bueh et al., 2008), in which the midlatitude anomaly center is related to the Bonin high. Enomoto et al. (2003) also suggested that the Bonin high is formed as a result of propagation of stationary Rossby waves along the Asian jet in the upper troposphere, which is called the Silk Road pattern. This can be clearly seen in the meridional wind field

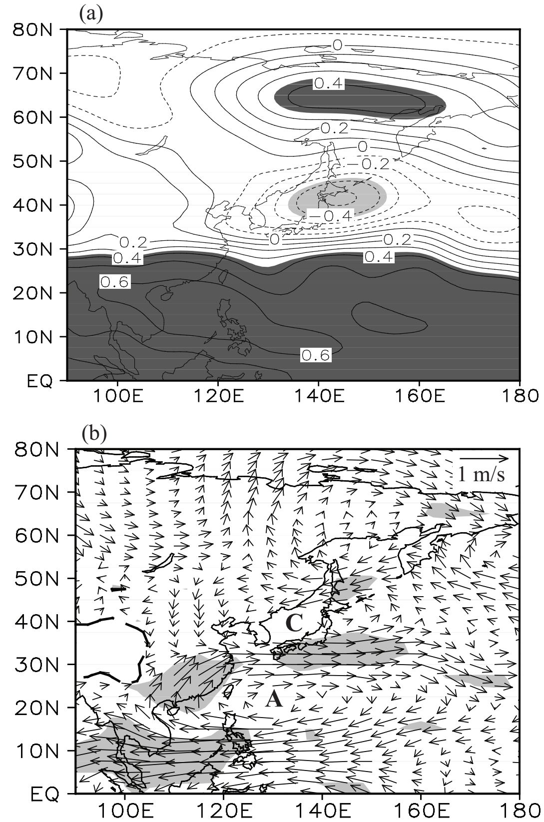

Fig. 4. The same as Fig. 3, but for the correlation (a) and regression (b) with the summer SCAP index.

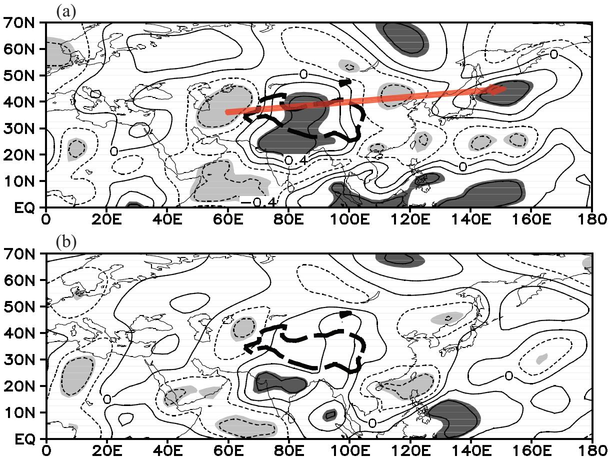

Fig. 5. Correlations of 200-hPa summer meridional winds with the (a) MB precipitation index and (b) SCAP index. The red arrow indicates the Silk Road pattern along the Asian jet, and the areas significant at the 95% confidence level are shaded.

at 200 hPa with alternative southerlies and northerlies along the Asian jet. Therefore, one may wonder whether the Silk Road pattern may modulate snow cover over the TP and alter the Bonin high and its associated MB precipitation at the same time. If that is the case, it implies that the effect of summer snow cover over the TP on MB precipitation may be doubtful. To test the contribution of the Silk Road pattern, we examine the correlations of 200-hPa summer meridional winds with the MB precipitation index (Fig. 5a) and with the SCAP index (Fig. 5b). In Fig. 5a, a positive–negative alternation pattern can be traced from Japan to the west of the TP along the Asian jet. This suggests that the MB precipitation is really connected with the Silk Road pattern. However, no clear positive–negative alternation can be identified in Fig. 5b. This indicates that summer snow cover anomalies over the TP cannot be attributed to the Silk Road pattern and the relationship between summer TP snow cover and MB precipitation cannot be explained by the simultaneous effects of the quasi-zonal propagation of stationary Rossby waves along the Asian jet. Therefore, the EAP pattern correlated with the SCAP index (Fig. 4a) is more likely induced by the meridional propagation of quasi-stationary planetary Rossby waves originating from anomalous convective activity in the western Pacific warm pool (Huang and Li, 1987; Nitta, 1987; Huang and Sun, 1992).

Next, we further explain the possible processes through which the SCAP affects the EAP pattern. The SCAP firstly modulates local surface temperature and heat source conditions over the TP. The simultaneous correlation between the summer SCAP and skin surface temperature (Fig. 6a) clearly illustrates a significant negative area over the western and southern TP. Meanwhile, in Fig. 6b, the total heat source (Q1) also shows negative correlations over the western and southern TP, although the significance level is relatively lower than in Fig. 6a. In addition, a significant negative correlation of Q1 appears over the Himalayas, which is a crucial area of summer atmospheric heat source (Duan and Wu, 2003). Here, the Q1 was calculated according to Zhao and Chen (2001). Further correlation analyses illustrate that the SCAP has stronger negative correlations with surface sensible heat (Fig. 6c) than with latent heat release of condensation (Fig. 6d). This implies that anomalous snow cover over the western and southern TP plays a more important role in modulating surface sensible heat and eventually affects the total heat source. The high (low) snow cover corresponds to low (high) surface temperature *in situ* and associated low (high) surface sensible heat and low (high) total heat source. This can be explained by the snow albedo and snow hydrological effects (Barnett et al., 1989). Through modulating the land surface temperature and atmospheric heat source, the SCAP may exert influences on atmospheric circulations.

Figure 6a also illustrates that, accompanying the negative correlation over the western TP, a large region of high positive correlation extends from the North Indian Ocean to the South China Sea. The out-of-phase correlation pattern implies that, corresponding to a colder TP (higher SCAP index), warmer surface temperatures appear in the North Indian Ocean region. A warmer North Indian Ocean can lead to anomalous upward motion, and vice versa. The simultaneous correlation of precipitation with the summer SCAP index (Fig. 7) shows that the North Indian Ocean region is occupied by a significant positive correlation, confirming the influence of ocean temperature on vertical motion. This result indicates that, accompanying the summer cold TP (the higher SCAP index),

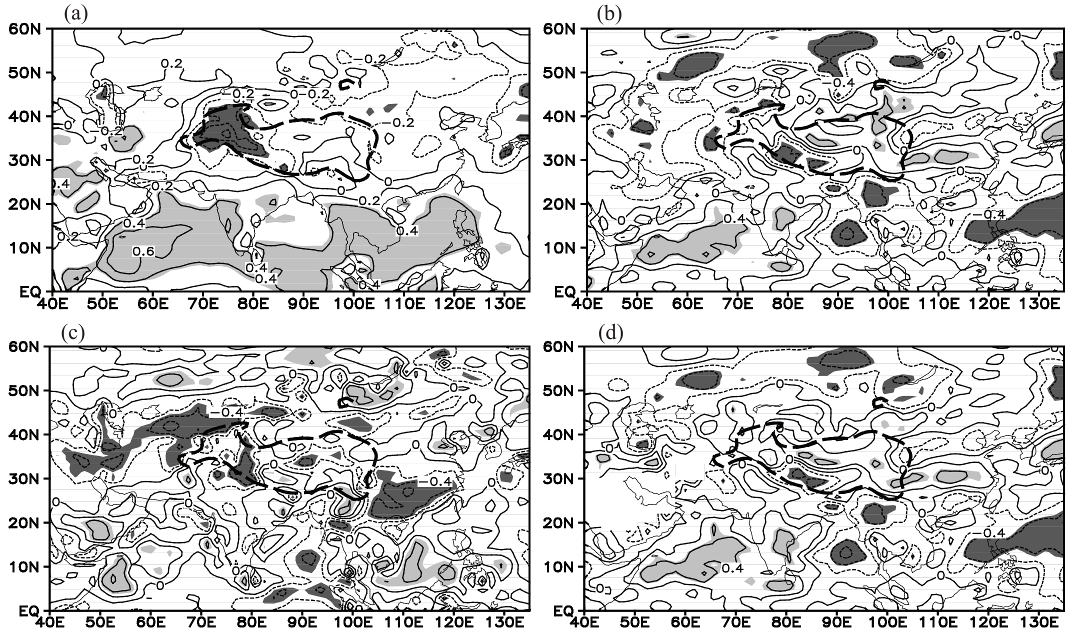

Fig. 6. Correlations of the summer SCAP index with (a) simultaneous skin temperature, (b) total heat source, (c) surface sensible heat, and (d) latent heat. The areas significant at the 95% confidence level are shaded.

anomalous upward motion may appear over the North Indian Ocean. Over the South China Sea–Philippine Sea, warmer surface temperature (Fig. 6a) corresponds to reduced surface heat fluxes (Figs. 6b–d) and suppressed precipitation (Fig. 7). Suppressed precipitation indicates more downward shortwave radiation. This correspondence signifies that the ocean surface temperature anomalies over the western North Pacific results from surface heat flux changes.

Figure 8a displays the regressed summer meridional vertical circulation along 80°E against the summer SCAP index. The longitude for the meridional vertical cross section was chosen based on Fig. 6. As shown in Fig. 8a, corresponding to higher SCAP index, air flow shows significant downward motion over the TP around 30°N, and compensatory upward motion around 10°–15°N. The anomalous upward motion around 10°–15°N plays an important role in increasing summer precipitation over the North Indian Ocean, as shown in Fig. 7. The above results imply that summer SCAP over the TP can to some extent modulate the vertical motion as well as related convection and precipitation over the North Indian Ocean.

The vertical motion over the North Indian Ocean modulated by the SCAP index can affect the WPSH through zonal vertical circulation in the tropics. The regressed summer zonal vertical circulation along 12.5°N against the summer SCAP index (Fig. 8b) shows that, accompanying a higher SCAP index, is notable anomalous upward motion around 80°E near the center of the North Indian Ocean anomalous precipitation, and then air flow moves eastward and sinks within 120°–140°E around the Philippines. The location of the anomalous downward motion generally corresponds to the center of anomalous geopotential height (Fig. 4a) and the western Pacific subtropical anticyclone (i.e. the WPSH, Fig. 4b). Wu et al. (2010) showed a similar east–west anomalous vertical circulation between the tropical Indian Ocean and the South China Sea–Philippine Sea in the interdecadal change around 1993. Figure 9 displays the regressed simultaneous 200-hPa divergent wind component and 500-hPa vertical p-velocity against the summer SCAP index, which further illustrates the processes linking the SCAP to summer rainfall along the MB band. Corresponding to a higher (lower) SCAP index, significant downward (upward) motion appears over the western TP, which is accompanied by upper-level (200-hPa) convergence (divergence) of anomalous divergent wind component in situ. Note that significant downward (upward) motion over the western TP related to the SCAP index corresponds to the positive (negative) correlation of precipitation in situ with the SCAP index (Fig. 7). This indicates that anomalous precipitation and vertical motion over the western TP do not match with each other, and this mismatch implies that anomalous western TP precipitation may not be primarily attributable to vertical motion over there. The vertical motion appearing over the

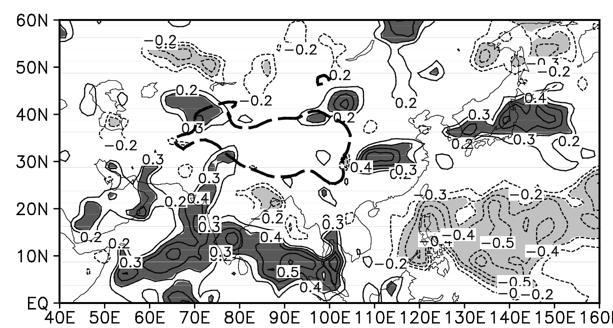

Fig. 7. Correlation between the summer SCAP index and simultaneous precipitation, with the shaded areas denoting correlation significant at the 90% confidence level.

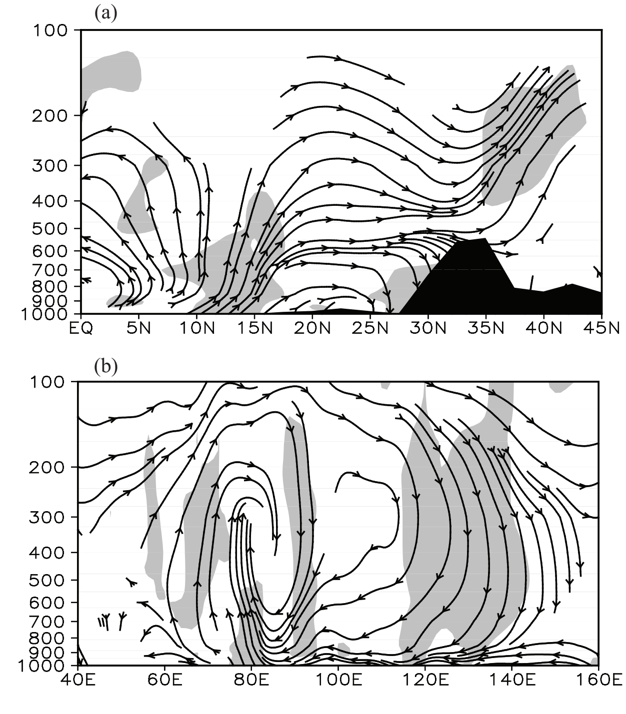

Fig. 8. Regressed simultaneous (a) meridional vertical circulation along 80°E and (b) zonal vertical circulation along 12.5°N against the summer SCAP index, with the shaded areas denoting significant vertical motion at the 95% confidence level.

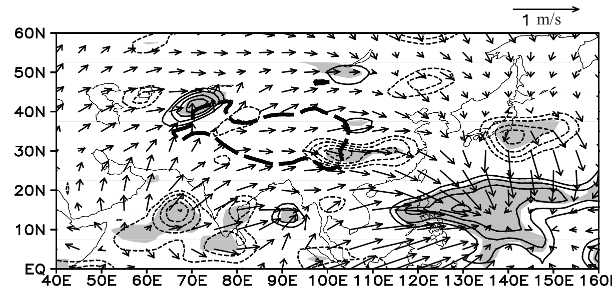

Fig. 9. Regressed simultaneous 200-hPa divergent wind component and 500-hPa vertical p-velocity (Pa s^{-1}, contours) against the summer SCAP index. The contour interval is 0.002 Pa s^{-1}. Shading denotes regions where the anomalies of regressed p-velocity are significant at the 95% confidence level.

western TP is linked with vertical motions over the Indian Ocean and the western Pacific through meridional and zonal vertical circulations, as shown in Fig. 8. Thus, in Fig. 9, significant upward (downward) motion and associated upper-level divergence (convergence) of anomalous divergent winds occur over the North Indian Ocean, and meanwhile significant downward (upward) motion and associated upper-level convergence (divergence) of anomalous divergent winds over the western Pacific warm pool. Eventually, this modulates the vertical motion over the MB region (Fig. 9). The result further demonstrates that the SCAP over the TP may remotely affect MB precipitation via the North Indian Ocean and the western North Pacific. Additionally, the increase in summer precipitation over the Indian Ocean could force a Kelvin wave response with low pressure and easterly wind anomalies over the equatorial Pacific, triggering a reduction in precipitation and the formation of a surface anticyclone over the subtropical Northwestern Pacific (Xie et al., 2009). This mechanism provides an alternative explanation as to how convective activity (upward motion) and precipitation over the North Indian Ocean modulated by the SCAP index affect the WPSH.

In summary, corresponding to a lower (higher) summer SCAP index, land surface temperatures and atmospheric heat sources in the western and southern TP are higher (lower). This leads to anomalous upward (downward) motion in situ and concurrently anomalous downward (upward) motion and associated suppression of convection and precipitation over the North Indian Ocean through a meridional vertical circulation from the TP to southern Asia and adjacent oceans. Meanwhile, through a zonal vertical circulation over the tropics, anomalous downward (upward) motion over the North Indian Ocean is closely linked to anomalous upward (downward) motion over the subtropical Northwestern Pacific, which directly results in a weaker (stronger) WPSH as well as associated stronger (weaker) convective activity and more (less) precipitation over the subtropical Northwestern Pacific (Fig. 7). On the other hand, an increase in convection over the North Indian Ocean can reinforce the WPSH through forcing a Kelvin wave response, which also explains the out-of-phase linkage between summer precipitation over the North Indian Ocean and that to the east of the Philippines (Fig. 7). In addition, summer anomalous convective activity in the western Pacific warm pool, which is closely related to the WPSH, should be considered as an important factor in stimulating the EAP (or PJ) pattern and accordingly affecting summer precipitation over the MB region (Nitta, 1987; Huang and Li, 1988; Huang and Sun, 1992). Therefore, through the above processes, the summer SCAP index is closely related to the EAP (PJ) pattern and associated summer precipitation over the MB region.

The upward motion and enhanced precipitation over the tropical Indian Ocean is essentially a response to the tropical Indian Ocean warming (Xie et al., 2009). The correlation between the summer MB precipitation index and simultaneous SSTs shows a large positive area (5°S–20°N, 50°–100°E) over the tropical Indian Ocean (not shown). This confirms that the SST anomalies in the tropical Indian Ocean may play an important role in modulating the vertical motion in situ and consequently affect the WPSH and associated MB precipitation. As such, one may wonder whether the summer SCAP and MB precipitation have no real physical linkage, and their close relationship is merely attributable to simultaneous impacts of the SST anomalies in the tropical Indian Ocean on the SCAP and MB precipitation. The summer SCAP is significantly associated with the simultaneous SST anomaly in the tropical Indian Ocean with a correlation coefficient of 0.48 at the 99% confidence level. In this respect, Fig. 7 does not definitively display the feature of an independent effect of the SCAP, which should therefore be further demonstrated.

The years (1980 and 1999) with more snow cover (SCAP index greater than 0.9σ) and the years (1988, 1990, and 2001) with less snow cover (SCAP index smaller than -0.9σ) but with weak tropical Indian Ocean SST anomalies (magnitude of the SST anomaly less than 0.5σ) were chosen to examine the independent effect of summer SCAP when the SST anomaly in the tropical Indian Ocean is relatively weak (as shown in Fig. 10a). In addition, the SST anomaly in the western Pacific warm pool is insignificant (Fig. 10a) and should not be considered as an important factor driving vertical motion and precipitation over the western Pacific warm pool. The composite difference in summer precipitation (Fig. 10b) shows significantly positive anomalies over India and the neighboring waters (primarily the eastern Indian Ocean) and over the MB region, and significantly negative anomalies over the western Pacific warm pool. The results confirm the independent effect of summer SCAP on upward motion and precipitation over the eastern Indian Ocean, which is a key component for meridional and zonal vertical circulations in Fig. 8. These results further demonstrate that summer SCAP in the TP is obviously an important factor and may to some extent independently modulate vertical motion over the Indian Ocean and the western Pacific warm pool, and eventually affect summer precipitation over the MB region.

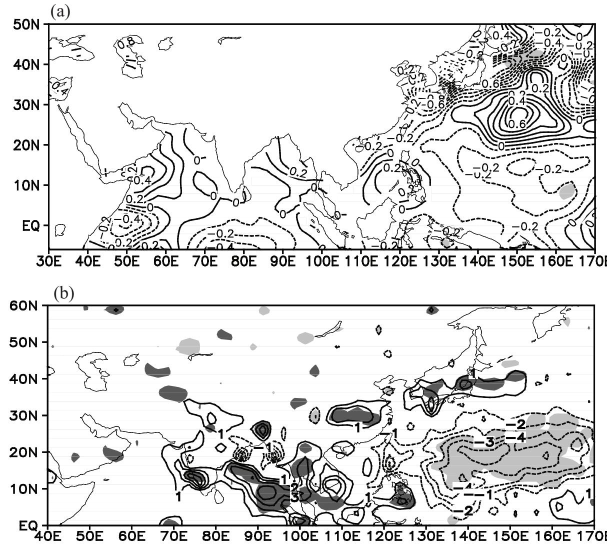

Fig. 10. Composite differences in (a) summer SSTs (°C) and (b) summer precipitation (mm) between the years (1980 and 1999) with more summer snow cover and those (1988, 1990, and 2001) with less summer snow cover when the SST anomalies in the tropical Indian Ocean are not clear. The shaded areas denote differences significant at the 90% confidence level.

5. Summary and discussion

In this study, we compared the abilities of several datasets in representing the summer snow anomaly in the TP and then chose the EASE-grid dataset to explore the relationship between the summer snow cover anomaly over the TP and simultaneous precipitation over East Asia as well as the potential mechanism underpinning this relationship.

Analysis revealed that although snow cover clearly shrinks in summer, it still exists in the western region and the southern flank of the TP where the altitude is high and accounts for approximately 23% of winter snow area, which can be identified in the satellite-derived EASE-grid dataset. However, the summer snow anomaly over the TP cannot be properly reflected by the 20th century reanalysis and station observations. The SCAP index, derived from the SWE and the percentage of visible snow of the EASE-grid dataset, appears more suitable for representing the variation in summer snow cover over the TP region.

The summer SCAP over the TP is significantly related to simultaneous precipitation over the MB region on the interannual time scale. Corresponding to higher (lower) SCAP over the TP, summer precipitation increases (decreases) over the MB region. Their close relationship can be explained by the following processes. Corresponding to lower (higher) summer SCAP index, land surface temperature in the western TP is higher (lower), which leads to anomalous upward (downward) motion over the western TP and simultaneously anomalous downward (upward) motion as well as associated suppression of convection and precipitation over the North Indian Ocean through meridional vertical circulation from the TP to the southern Asian adjacent oceans. Through zonal vertical circulation in the tropics and the Kelvin wave response processes, anomalous downward (upward) motion in the North Indian Ocean may give rise to anomalous upward (downward) motion over the subtropical Northwestern Pacific, and modulate the WPSH and associated convective activity in the western Pacific warm pool. Finally, anomalous convective activity in the western Pacific warm pool stimulates the EAP (or PJ) pattern and consequently affects summer precipitation over the MB region through the meridional propagation of quasi-stationary planetary Rossby waves. Although it plays an important role in the formation of the EAP pattern (Enomoto et al., 2003; Bueh et al., 2008), the zonal propagation of stationary Rossby waves along the Asian jet seems not to be responsible for the relationship between the SCAP and anomalous EAP pattern.

Typical years with stronger snow cover anomalies and with no clear SST anomalies in the tropical Indian Ocean at the same time were examined to investigate the independent effect of the summer SCAP. Composite analysis revealed that the independent effect of the summer SCAP may modulate vertical motion over the North Indian Ocean, particularly over its eastern part, and eventually affect summer precipitation over the MB region through the above-mentioned mechanism.

The SCAP index shows a good persistence in the interannual variation, especially from the previous May to summer, with a high correlation coefficient of 0.50 at the 99% confidence level (Liu et al., 2014). Because of the high persistence of the SCAP index from May to summer, the May SCAP index is significantly related to summer precipitation over the MB region (a correlation coefficient of 0.51 at the 99% confidence level) and could be regarded as a forecast factor. The persistence of the SCAP index can partly explain the season-delayed effect of snow cover over the TP on summer rainfall over the MB region, which is different from the mechanism of a soil moisture bridge suggested by Zhao et al. (2007). Moreover, the path of the impact of the SCAP on summer precipitation over East Asia, which is through the North Indian Ocean and the western Pacific warm pool, is also different from the path via the south Asian jet stream waveguide in the impact of spring snow cover over the eastern TP in the previous study (Zhao et al., 2007). Therefore, the effect of SCAP in the present study should be considered as a supplement to previous work

Additionally, it should be stated that the upward motion (convective activity) over the North Indian Ocean, which is a key factor in modulating the WPSH and summer precipitation over the MB region, is obviously not just affected by the SST anomaly in the tropical Indian Ocean. The SCAP over the TP may also exert a crucial influence and should not be ignored. Moreover, the correlation of tropical central-eastern Pacific SST with the summer SCAP index (not shown) is weak in winter and spring and significantly increases in summer. This differs from the correlation with summer precipitation over the MB region, which is very significant in winter and spring and clearly decreases in summer (not shown). This implies that tropical central-eastern Pacific SST anomalies have different connections with summer snow cover over the TP and summer MB precipitation, and the close relationship between the summer SCAP and summer MB precipitation cannot be simply attributed to the simultaneous effects of SST anomalies in the tropical central-eastern Pacific. Although the potential mechanism for the effect of SCAP has been investigated using statistical analyses, the detailed processes of its impact on summer precipitation over East Asia remain unclear and should be further explored by simulation experiments. Forecasting summer precipitation over the MB region by synthetically using the preceding SST signals and the preceding SCAP signal over the TP is also worthy of further exploration.

Acknowledgements. The authors appreciate the comments and suggestions of the two anonymous reviewers, which greatly improved this paper. We also thank the National Snow and Ice Data Center (NSIDC), the National Centers for Environmental Prediction (NCEP), the Physical Sciences Division (PSD) of the National Oceanic and Atmospheric Administration (NOAA), and the National Climate Center (NCC), China Meteorological Administration, for providing data via their websites. YZ acknowledges the support of the National Natural Science Foundation of China (Grant No. 41271434), the National Key Technologies R&D Program in the 12th Five Year Plan of China (Grant No. 2012BAH32B03), the Hong Kong GRF (Grant No. CUHK 457212), and the ITF (Grant No. GHP/002/11GD). RW acknowledges the support of the Hong Kong Research Grants Council (Grant No. CUHK 403612) and the National Natural Science Foundation of China (Grant Nos. 41275081 and 41228006). GL acknowledges the support of the National Natural Science Foundation of China (Grant Nos. 41375090 and 41221064), the Special Project of the National International Science and Technology Cooperation of China (Grant No. 2011DFG23450), and the Basic Research Fund of CAMS (Grant No. 2013Z002).

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