Impacts of ENSO and IOD on Snow Depth Over the Tibetan Plateau: Roles of Convections Over the Western North Pacific and Indian Ocean

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Abstract

There is a consensus that snow over the Tibetan Plateau (TP) modulates the regional climate significantly. Possible causes for the interannual variability of snow over the TP, however, are under debate, especially regarding the independent roles of El Niño-Southern Oscillation (ENSO) and Indian Ocean dipole (IOD). Based on in situ observational data analyses and model simulations, our study shows that impacts of ENSO and IOD on snow depth (SD) over the TP are different during early winter. In particular, ENSO mostly affects SD over the eastern TP, while IOD affects SD over the central western TP. Both above-normal snowfall and cold temperature anomaly contribute to deeper-than-normal SD, with the former playing a more important role. Diabatic cooling of the suppressed convection over the western North Pacific that related to the positive phase of ENSO could excite an anomalous cyclonic circulation and strong cold temperature anomalies over the eastern TP. There is an enhanced moisture transported over the eastern TP from the tropics due to the anomalous cyclonic circulation, along with strong cold temperature anomalies, resulting in above-normal snowfall in the eastern TP. On the other hand, anomalous convection over the western Indian Ocean related to the positive IOD could generate a wave train propagating northeastward and induce an anomalous cyclonic circulation over the central western TP. The associated anomalous circulation transports extra moisture from the tropics to the central western TP, providing conditions favorable for more snowfall over the central western TP. Opposite conditions tend to occur during negative phases of ENSO and IOD.

1. Introduction

The Tibetan Plateau (TP), located over the central and eastern parts of the Eurasian continent, is the largest and highest plateau in the world. The multiscale mountain ranges have significant impacts on atmospheric circulation, energy, and water cycles of the climate system by dynamics and thermodynamic effects (e.g., Boos & Kuang, 2010; Boos & Kuang, 2013; Yanai et al., 1992; Wu & Zhang, 1998; Wu & Liu, 2003; Wu, Liu et al., 2012). Earlier theoretical and numerical modeling studies have mainly focused on the mechanical forcing (Bolin, 1950; Charney & Eliassen, 1949; Yeh, 1950), while more and more attention has been paid to the thermal forcing of TP during recent decades (Yanai et al., 1992; Wu & Zhang, 1998; Duan & Wu, 2005; Wu et al., 2007, 2012; Jiang et al., 2013, 2015, 2016; Wang et al., 2019).

Snowpack exerts great influence on the thermal status over the TP and thus plays important roles in weather and climate over and beyond Asia (Bamzai & Shukla, 1999; Wang et al., 2017; Xie et al., 2005; Yeh et al., 1983; Zhao & Chen, 2001). Increase of snowpack tends to reduce solar radiation and consume extra solar energy for snow melting, thus resulting in an anomalous lower land surface heating and overlying air temperature (Shukla & Mooley, 1987; Yasunari et al., 1991). The winter and spring snow anomaly effect could persist to the subsequent summer (Shaman & Tziperman, 2005; Yuan 2012; Zhang et al., 2019) and thus links to the anomalies of East Asian summer monsoon and Indian summer monsoon (Ding et al., 2009; Fasullo, 2004; Wu & Qian, 2003). The winter and spring snow anomaly can also affect typhoon activities.
over the western Pacific (Wu & Qian, 2003; Xie et al., 2005; Zhang et al., 2004). In addition to winter and spring, snow anomaly over the TP in summer affects climate over East Asia and Europe (Wu et al., 2012; Wu et al., 2016). Therefore, understanding of the TP snow variation provides insights into the variation and prediction of climate and weather worldwide.

Some studies have investigated the variability of snow over the TP and possible causes. Using the Nimbus-7 Scanning Multichannel Microwave Radiometer (SMMR) satellite estimation of snow depth from November 1978 to August 1987, Shaman and Tziperman (2005) proposed that an El Niño might excite the stationary Rossby waves extending along the North African-Asian jet, resulting in anomalous increase of potential vorticity and snow depth over the TP in winter. This mechanism was confirmed by Wang and Xu (2018), who used snow water equivalent data from 1987 to 2007. On the other hand, Yuan et al. (2009, 2012) reported that the interannual variability of the winter TP snow cover is linked to Indian Ocean dipole (IOD) rather than El Niño-Southern Oscillation (ENSO). They demonstrated that in the pure ENSO years with no co-occurrences of the IOD, the influence of ENSO on the early winter TP snow cover is negligible (Yuan et al., 2009). Using reanalysis data set from ERA-Interim, Zhang et al. (2019) also emphasized on IOD forcing on snow depth over the TP. It is interesting why these studies have obtained different conclusions regarding the impacts of ENSO and IOD on snow over the TP. Of note is that two satellite-based or reanalysis-based snow parameters have been investigated in these studies: snow cover and snow depth (Shaman & Tziperman, 2005; Wang & Xu, 2018; Yuan et al., 2009, 2012; Zhang et al., 2019). The snow depth derived from satellite show considerable biases (Frei et al., 2012). Thus, it is necessary to use in situ snow depth data to reinvestigate the relationship of snow depth over the TP with ENSO and IOD.

While Shaman and Tziperman (2005) proposed that ENSO affects the snow depth over the TP by exciting a stationary Rossby wave extending along the North African-Asian jet, many previous studies have reported that ENSO could affect East Asian climate by modulating convection over the western Pacific (Chang et al., 2000; Lau & Nath, 2006; Wang et al., 2000). Jiang et al. (2013) found that interannual variability of wintertime surface air temperature over the southeastern TP is linked to convection anomaly over the western North Pacific (WNP), which is significantly affected by ENSO. As surface air temperature can affect snow depth/snow cover (Bao & You, 2019; Bojariu & Gimeno, 2003; Wang et al., 2019), is it possible that ENSO can affect snow depth/snow cover over the TP by modulating the convection over the WNP?

Compared to snow cover, snow depth is better at expressing the capacity of snow affecting the overlying airflow and has a more significant connection to summer monsoon rainfall (Barnett et al., 1989; Kripalani & Kulkarni, 1999; Ye & Bao, 2001). Thus, this study aims to answer whether ENSO and IOD can have an impact on snow depth over the TP. If yes, what are the independent roles of ENSO and IOD on TP snow depth anomalies? Can ENSO (IOD) affect snow depth over the TP by modulating convection over the WNP (Indian Ocean [IO])? This paper is arranged as follows. In section 2, the observational data, the analysis method, and the model experiments are described. In section 3, temporal and spatial variabilities of the TP snow depth and its relationship with ENSO and IOD are depicted. In section 4, different impacts of ENSO and IOD on snow depth over the TP and responsible mechanisms are discussed. Finally, a summary of the study and further discussions are provided in section 5.

2. Data, Method, and Model

2.1. Data

Observed daily data sets of snow depth, snowfall, and 2-m air temperature (T2M) for the period of 1980–2010 in China, compiled by the China Meteorological Administration after quality control, are used in this study. It should be noted that the snow depth is defined as the depth of snow layer in the observation site. There are 2,481 stations for snow depth and T2M observation and 836 stations for snowfall observation. Snowfall is identified using the weather phenomena record, following Zhou et al. (2018). Further quality control was conducted for all station observations as follows: (i) missing days of each month should be fewer than 14 days; (ii) no missing month is allowed; (iii) if a station was relocated, the horizontal distance and differences of elevation between the new site and the old site should be less than 10 km and 100 m, respectively; and (iv) observations for snow depth, snowfall, and T2M should be available for a particular station. Here we selected
stations with elevation above 1,000 m within the TP, so that totally 79 stations are considered in this study (Figure 1).

Other data sets including monthly temperature, winds, and moisture flux from ERA-Interim with a resolution of 0.75° × 0.75° (Dee et al., 2011); monthly sea surface temperature (SST) from the Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST) version 1.1 with a resolution of 1° × 1° (Rayner et al., 2003); and monthly National Oceanic and Atmospheric Administration (NOAA) Interpolated Outgoing Longwave Radiation (OLR) with resolution of 2.5° × 2.5° (Liebmann and Smith, 1996) are also used in this study.

2.2. Method

Partial correlation analysis

\[ R_{ab,c} = \frac{(R_{ab} - R_{ac}R_{bc})}{\sqrt{(1-R_{ac}^2)(1-R_{bc}^2)}} = \frac{(R_{ab} - R_{ac}R_{bc})}{\sqrt{(1-R_{ac}^2)}} \]

is applied to exclude the possible influence by ENSO or IOD, where \( R_{ab,c} \) is the partial correlation coefficient between variable \( a \) and variable \( b \) without influence of variable \( c \). In addition, linear regression and multiple linear regression are applied in this study. The linear regression is calculated by \( y = bx + a \), where \( b \) is the regression coefficient. The multiple linear regression is calculated by \( y = b_1x_1 + b_2x_2 + a \), where \( b_1 \) and \( b_2 \) are the partial regression coefficients for \( x_1 \) and \( x_2 \), respectively. All time series used for computing correlation in this study are linearly detrended. In this study, the Niño-3, Niño-3.4, and Niño-4 indices are

Figure 1. Climatological (1980–2010) snow depth (units: cm) over the TP from August to July.
defined as the SST anomalies averaged over the areas of 5°S–5°N, 90–150°W; 5°S–5°N, 120–170°W; and 5°S–5°N, 160°E–150°W, respectively. The IOD index is defined as the SST difference between the western equatorial IO (10°S–10°N, 50–70°E) and the eastern equatorial IO (10°S–0°, 90–110°E) (Saji et al., 1999).

2.3. Model
To investigate the impacts of diabatic heating due to convection anomalies on atmospheric circulation, a nonlinear baroclinic model, developed by Ting and Yu (1998), is employed. This model is a fully nonlinear, dry, time-dependent baroclinic model with 24 sigma levels in the vertical and spectral R30 horizontal resolution. Climatological zonally varying basic state for this model is taken from the ERA-Interim reanalysis (Dee et al., 2011). Dissipations employed in the model include Rayleigh friction, Newtonian cooling, and biharmonic diffusion (coefficient of $1 \times 10^{17} \text{ m}^{-4} \text{ s}^{-1}$). The coefficients of Rayleigh friction and Newtonian cooling are the same, with timescales of 0.3, 0.5, 1.0, and 8.0 days in the lowest four levels and 15 days in the other levels. The time integration is performed for 50 days. The model solution approaches a steady state after about 15 days, and the average for the last 20 days are shown in this study.

The idealized heating is prescribed as $Q = V(\sigma)A(\lambda, \phi)$. The vertical structure of the heating takes the form $V(\sigma) = e^{-20(\sigma-\sigma_c)^2}$. It has a maximum when $\sigma$ equals $\sigma_c$ and reduces to zero quickly as $\sigma$ increases or decreases from $\sigma_c$. $\sigma_c$ is chosen to be 0.37. The function $A(\lambda, \phi)$ represents the horizontal structure and magnitude of the heating.

3. Temporal and Spatial Variabilities of the TP Snow Depth and Its Relationship With ENSO and IOD
Figure 1 shows climatology of snow depth over the TP from August to July. Snow depth starts to build up in September, especially over the central TP (Figure 1b). Snow depth greater than 0.5 cm appears first over the central eastern TP in October (Figure 1c) and then over the western TP in November (Figure 1d). It continues to increase in the subsequent months (Figures 1c–1f) and peaks in January, with values greater than 1 cm over the entire TP. It then starts to decay and generally less than 0.5 cm over the western TP in April, while it is around 0.5–2 cm over the eastern TP (ETP) (Figure 1i). Snow depth continues to decrease in May and June (Figures 1j and 1k). Note that snow nearly disappears during July–August, which is different with features revealed by the satellite observation (Yuan et al., 2009) (Figure 1). This inconsistency may be due to the fact that most of the stations with in situ snow depth measurements are located in the valley.

Figure 2 presents annual cycles of snow depth, snowfall, and 2-m temperature averaged over the TP. It can be seen that both the decrease in T2M and the increase in snowfall before January contribute to the snow depth maximum in January. After that, snowfall increases and reaches a maximum in March and then decreases gradually. On the other hand, T2M increases after January. Thus, the decrease in snow depth after January is mostly caused by the increase in T2M. Thus, the process related to variation of the snow depth is somewhat different between after and before January.

Previous studies reported that interannual variability of snow over the TP is linked to ENSO and IOD (e.g., Shaman & Tziperman, 2005; Yuan et al., 2009). To measure the variability of snow depth, the TP snow depth index (TSDI) is defined as the regional-averaged snow depth over the entire TP. Figure 3 shows correlation of the TPSDI with ENSO and IOD for each calendar month. Significant positive correlation between TP snow depth and ENSO in September and between ENSO and also IOD in December is found (Figure 3). The lead-lag correlation between the TPSDI and the

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**Figure 2. Annual cycles of the snow depth (cm; black solid line), snowfall (mm day$^{-1}$; black dashed line), and 2-m temperature (°C; red solid line) averaged over the TP.**

**Figure 3. Coefficients of correlations between the TPSDI and ENSO/IOD indices from August to July. The dashed line denotes the correlation coefficients at a 90% confidence level by a t test.**
ENS/O/IOD index shows that the December TP snow depth is significantly correlated with ENSO (IOD) 3 months (1 month) in advance (figure not shown). Insignificant relationship between the TP snow depth and ENSO/IOD can be seen during the other months (Figure 3). In regard to spatial features, significant positive correlation of snow depth with ENSO and IOD is found over the entire TP region in November and December and over the southwestern TP in October (figures not shown). However, the variability of the TPSDI is dominated by snow depth over the central TP (where snow depth is greatest) in October (figure not shown), thus contributing to a low correlation between the TPSDI and ENSO/IOD as shown in Figure 3. From Figure 3, it is also noticed that the correlation values from August to December are different from those in the rest of the months. The correlation coefficients decrease significantly from December to January and remain small or even negative (Figure 3). This shift might be ascribed to different contributions of T2M and snowfall to snow depth as revealed by Figure 2. Since considerable persistent positive correlation between the TPSDI and the ENSO/IOD index only appears in November and December, we focus on the influence of ENSO/IOD on TP snow depth during November–December in this study. Figure 4 shows the normalized TPSDI, Niño-3 index, and IOD index during November–December. The interannual variability of the TPSDI matches well with the interannual variability of the Niño-3 index and IOD index, particularly in strong ENSO years (e.g., 1997). Correlation coefficients of the TPSDI with the Niño-3 and IOD indices are 0.52 and 0.61, respectively, both exceeding 99% confidence level.

Correlation maps of snow depth with the Niño-3 and IOD indices during November–December are shown in Figures 5a and 5b, respectively. Significant positive correlation between snow depth and the Niño-3 index appears over the entire TP (Figure 5a). In particular, values exceeding 99% confidence level are mostly located over the ETP (Figure 5a). In comparison, significant positive correlation between the snow depth and the IOD index tends to appear more over the central western TP (CWTP) than over the ETP region (Figure 5b). To measure variability of snow depth over the CWTP and the ETP, we define the CWTP (ETP) snow depth index (CWTPSDI (ETPSDI)) as regional-averaged snow depth over the CWTP (ETP). The ETP (CWTP) is defined as east (west) of 98°E within the TP region. Correlation coefficients of the Niño-3 index with the ETPSDI and the CWTPSDI are 0.512 and 0.345, respectively (Table 1). On the other hand, there is significant positive correlation between the IOD index and the CWTPSDI (0.642) but insignificant correlation between the IOD index and the ETPSDI (0.287) (Table 1).

The regional features of the relationship between ENSO/IOD and snow depth over the TP can be seen more clearly in Figures 6a and 6b, which show partial correlation of snow depth with the Niño-3 and IOD indices during November–December. After excluding the influence from IOD (ENSO), significant correlation between snow depth and the Niño-3 (IOD) index is only seen over the ETP (CWTP) (Figures 6a and 6b). Partial correlation coefficients of Niño-3 (IOD) index with the ETPSDI and the CWTPSDI are 0.455 (−0.118) and −0.193 (0.598), respectively (Table 1). The above suggests that influence of ENSO on snow depth over the CWTP during early winter may be mainly through its modulation on IOD. On the other hand, there is an independent influence of ENSO on snow depth over the ETP during early winter. This result is distinctive from that in Yuan et al. (2009), who indicated a negligible influence of ENSO on the early winter TP snow cover with no co-occurrences of IOD, by using satellite-observed snow cover data.

4. Mechanisms

4.1. Roles of Local Snowfall and Surface Air Temperature

As discussed in section 3, the TP snow depth is directly affected by both surface air temperature and snowfall climatologically before January. Here we discuss the possible relationships of surface air temperature and snowfall with snow depth on interannual time scale. Figure 7 shows correlations of snow depth with local snowfall and T2M during November–December. Interannual variability of snow depth is positively and significantly correlated with snowfall over the entire TP, while correlation between snow depth and T2M is negative and relatively lower than that between snow depth and snowfall, especially over the ETP region.
This suggests that snowfall has a greater influence than has surface air temperature on snow depth across the TP in early winter.

Correlation analysis reveals that ENSO and IOD have different impacts on snow depth over the ETP and the CWTP. Figures 5c–f and 6c–f show correlation and partial correlation between the Niño-3/IOD index and snowfall/T2M during November–December. The spatial patterns for snowfall are largely consistent with those for snow depth (Figures 5 and 6). There is also a similarity between spatial patterns for T2M and snow

Table 1
Correlation (Partial Correlation) of the Niño-3/IOD Index with the CWTPSDI/ETPSDI During November–December

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*Coefficient exceeding the 90% confidence level. **Coefficient exceeding the 95% confidence level. ***Coefficient exceeding the 99% confidence level. ****Coefficient exceeding the 99.9% confidence level.

Figure 5. Correlations of (a–b) snow depth, (c–d) snowfall, and (e–f) T2M with the (left) Niño-3 index and (right) IOD index during November–December. Green dashed line denotes boundary of the CWTP and ETP. Values of ±0.301, ±0.355, and ±0.456 represent 90%, 95%, and 99% confidence levels, respectively.

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Figure 6. Partial correlations of (a–b) snow depth, (c–d) snowfall, and (e–f) T2M with the (left) Niño-3 index and (right) IOD index during November–December. Green dashed line denotes boundary of the CWTP and ETP. Values of ±0.306, ±0.361, and ±0.463 represent 90%, 95%, and 99% confidence levels, respectively.

Figure 7. Correlations of snow depth with (a) snowfall and (b) T2M during November–December. Values of ±0.301, ±0.355, and ±0.456 represent 90%, 95%, and 99% confidence levels, respectively.
depth, but lower correlation coefficients between T2M and the Niño-3/IOD index are shown over the TP (Figures 5 and 6). These features again indicate that ENSO (IOD) has a closer link to climate anomalies over the ETP (CWTP). The values of correlation coefficients suggest that the independent influence of ENSO (IOD) on snow depth over the ETP (CWTP) may be mainly through the modulation on snowfall, while the modulation on T2M is of secondary importance (Figures 5 and 6).

4.2. Atmospheric Circulation, Temperature, and Convection Anomalies

ENSO and IOD exert impacts on remote climate by modulating tropical convection and associated large-scale atmospheric circulation and temperature anomalies (e.g., Lee et al., 2017; McPhaden et al., 2006; Sohn et al., 2016; Wang et al., 2000; Yuan et al., 2009, 2012; Zhang et al., 2016). Here we examine anomalies of large-scale circulation, temperature, and convection associated with interannual variability of snow depth over the TP. As the snow depth over the CWTP and the ETP have different links to IOD and ENSO, we calculate the large-scale circulation, temperature, and convection anomalies related to snow depth variations over the CWTP and ETP, respectively, and results are shown in Figure 8.

Figure 8 (left column) shows anomalies related to interannual variability of the CWTP snow depth. In the lower troposphere, there are positive OLR anomalies over the western Maritime Continent (WMC) and negative OLR anomalies over the western tropical IO (WIO) (Figure 8a). Correspondingly, there is anomalous easterly flow over the Equator from the WMC to the WIO, accompanied by anomalous anticyclonic circulations over the southern tropical IO and which stretches from the Arabian Peninsula to the WNP.
There is anomalous southwesterly flow from the tropical sea to the CWTP (Figure 8a). In the middle and upper troposphere, warm temperature anomaly appears over northern Africa and WIO, while cold anomaly is found over northern India and the CWTP. There is a wave train that propagates northeastward along the South Asian waveguide (Figures 8c and 8e). It induces barotropic cyclonic circulation anomalies over northern India and the CWTP. The anomalous westerlies turn cyclonically from northern India toward the CWTP.

Atmospheric anomalies related to interannual variability of snow depth over the ETP are shown in the right panels of Figure 8. In the lower troposphere, there are positive OLR anomalies over the WNP and negative OLR anomalies over the WIO (Figure 8b). There is an anomalous easterly flow over the equatorial IO, and a pair of anomalous anticyclonic circulation straddling the Equator, similar to those for the CWTP snow depth to some extent (Figure 8b). The most notable difference is that there is a strong anomalous cyclonic flow to the east of the TP, contributed by a pair of anomalous anticyclonic circulation over the WNP and across Japan (Figure 8b). In the middle and upper troposphere, anomalous cyclonic circulation and cold temperature anomaly are dominant across the TP (Figures 8d and 8f). Different with that for the CWTP snow depth, westerly anomalies in the southern flanks of the anomalous cyclonic circulation do not turn into southwesterly anomalies in the southern TP, and circulation anomalies over the WNP are different between 500 and 300 hPa. However, there is a cyclonic circulation anomaly to the southeast of TP in both 500 and 300 hPa, accompanied by southeasterly anomalies from the WNP toward the eastern edge of TP (Figures 8d and 8f). There is a significant anomalous cold center from the ETP to the WNP at 300 hPa.

Figure 9 shows regressions of vertically integrated (from 500 to 200 hPa) moisture flux and its divergence against the CWTP and ETP snowfall indices (defined as regional-averaged snowfall over the CWTP and ETP), respectively. The increase in snowfall over the CWTP is accompanied by a cyclonic moisture flux pattern over the TP, with southwesterly moisture transport at the southern boundary. In general, there is convergence of water vapor flux over most of the TP. Comparison between Figures 8c, 8e, and 9a indicates that the cyclonic circulation anomalies associated with the wave train could transport moisture from the tropical sea to the CWTP. Therefore, the extra moisture supply, together with a colder air temperature, provide a favorable condition for enhancement of snowfall and hence deepening of snow depth over the CWTP (Figures 8c, 8e, and 9a). On the other hand, the increase of snowfall over the ETP is corresponding to surplus moisture transport from the WNP to the ETP by anomalous easterlies and westerlies, respectively, resulting in moisture convergence over the ETP, in particular the southeastern TP (Figure 9b). The anomalous easterlies from the WNP to the ETP is associated with an anomalous cyclonic circulation to the southeast of the TP. Obviously, the atmospheric conditions depicted above (i.e., moisture convergence over the ETP and cold temperature anomalies over the TP) are favorable for accumulation of snowfall and deepening of snow depth over the ETP.

To disclose the mechanisms responsible for different impacts of ENSO and IOD on snow depth over the CWTP and the ETP, partial regressions of OLR, temperature, and winds against the Niño-3/IOD index are further calculated (Figure 10). As shown in Figure 10 (left column), the partial regression patterns at different levels with regard to the Niño-3 index after exclusion of the influence of IOD show a large similarity with those for the ETPSDI, including the OLR anomalies over the WNP, the anomalous anticyclonic circulation
over the WNP at low level, the anomalous cyclonic circulation to the southeast of the TP at 500 and 300 hPa, and the anomalous cold center at 300 hPa from the ETP to the WNP. Notable differences are that the anomalous cyclonic circulation to the southeast of the TP becomes quite strong in the upper troposphere, but the anomalous cyclonic circulation and temperature anomalies to the west of the TP are relatively weaker (Figures 10c and 10e). From Figure 11a, which shows the partial regressions of vertically integrated (300 to 500 hPa) moisture flux and its divergence onto the Niño-3 index, there is in general moisture convergence over the ETP, with a maximum center over the southeastern TP, consistent with features in Figure 9b. Comparisons of moisture flux and temperature anomalies associated with the Niño-3 index, the ETP snowfall, and snow depth indicate that the anomalous cyclonic water transport to the southeast of the TP and the temperature anomalies from the ETP to the WNP are important for above-normal snowfall and deepening of snow depth over the ETP.

Figure 10 (right column) shows partial regression patterns at different levels with regard to the IOD index after the influence of ENSO is excluded, which bears great resemblance with those for the CWTPSDI (Figure 8, left column), including OLR anomalies over the WMC and WIO. Compared to features for the CWTP snow depth, the anomalous warming shows a northward shift to Arabian Peninsula-WIO region in the middle troposphere and becomes even stronger in the upper troposphere (Figures 10d and 10f). This anomalous warming excites a barotropic Rossby wave with northeastward dispersion along the South Asian wave guide in both middle and upper troposphere (Figures 10d and 10f), consistent with features in

Figure 10. (top row) Partial regressions of OLR (W m$^{-2}$) and 850-hPa winds (m s$^{-1}$) against (left) the Niño-3 index and (right) the IOD index. Regressions of air temperature (K) and winds (ms$^{-1}$) at (middle row) 500 hPa and (bottom row) 300 hPa against (left) the Niño-3 index and (right) the IOD index. Significant values exceeding the 90% confidence level are marked by purple dots or black vectors.
Yuan et al. (2009, 2012). Although there is an insignificant temperature anomaly over the CWTP in the middle and upper troposphere, the anomalous cyclonic circulation over the northern India-CWTP induced by the wave train favors moisture transport from the northern IO toward the CWTP, thus benefiting snowfall over the region (Figures 10d, 10f, and 11b).

4.3. Role of Tropical Convections

From above analyses, ENSO (IOD) has an impact on snow depth over the ETP (WTP), mainly through its modulation on atmospheric circulation and temperature over the TP and surrounding regions. The atmospheric circulation anomalies, according to the Gill (1980) model, may be explained as a response to diabatic heating related to convection anomalies over the tropical oceans as shown in Figures 8 and 10. It is known that convection over the WNP (WIO and WMC) is closely linked to ENSO (IOD), so is it possible that convection over the WNP (WIO and WMC) works as a medium by which ENSO (IOD) affects the ETP (WTP) snow depth?

In this section, we investigate the individual and combined impacts of diabatic heating of the three convection anomalies on regional atmospheric circulation. As convection over the three regions is highly correlated to each other in observation, it is difficult to identify individual impact of convection over one of the three regions based on observation data. Thus, here we use a nonlinear baroclinic model to simulate the individual and combined impacts of convection anomalies over the three regions on regional circulation and temperature anomalies. An introduction of the model has been given in section 2.3. In the vertical, the heating has a sinusoidal profile with a maximum at sigma of 0.37 to mimic the condensational heat released from deep convection (Figure 12a). The horizontal structure and magnitude of the heating are dependent on the regressions of rainfall from GPCP against the Niño-3 index and IOD index, respectively (Figure 12b). The maximum value of heating rate with the unit of Kelvin per day equals the double value of the regression coefficient of rainfall amount with the unit of millimeter per day (Figure 10). Several experiments are conducted: Heating is added over (1) the three regions of the WIO (15°S–10°N, 20°–60°E), WMC (15°S–5°N, 80°–110°E), and WNP (5°–20°N, 105°–150°E) (Figure 12b); (2) the two regions of the WIO and WMC; (3) the region of the WNP; (4) the region of the WIO; and (5) the region of the WMC. Heating is limited to the region where regression coefficient exceeds 90% confidence level.

Figure 11. Partial regressions of vertically integrated from 500- to 200-hPa moisture flux (kg m$^{-1}$ s$^{-1}$; vectors) and its divergence (10$^{-6}$ kg m$^{-2}$ s$^{-1}$; shading) against (a) the Niño-3 index and (b) the IOD index. Significant values exceeding the 90% confidence level are marked by purple dots or black vectors.

Figure 12. (a) Vertical profile of specific heat source (K day$^{-1}$) around the horizontal maximum heating center, (b) spatial pattern of specific heat sources (shading; K day$^{-1}$) at the level of sigma that equals 0.37. The x-axis and y-axis in (a) denote the magnitude of the heating and height represented by sigma, respectively.
Combined forcings over the three regions excite easterly flow over the equatorial IO and a pair of anticyclonic circulation to its flanks (Figure 13a). There is anticyclonic circulation over the Arabian Peninsula, and cyclonic circulation over the northern Indian subcontinent, similar to the wave train associated with IOD. On the other hand, there is strong cyclonic circulation from the southeastern TP to the WNP, accompanied by strong negative temperature. The spatial patterns are similar to patterns of wind and temperature regression onto the TPSDI (figure not shown), indicating that the convection anomalies associated with ENSO and IOD forced circulation anomalies favor for deepening of snow depth over the whole TP (Figures 10 and 13a).

As a Gill-type response to combined heating over the WIO and WMC, there is easterly flow over the equatorial IO and a pair of anticyclonic circulation to its flanks (Gill, 1980) (Figure 13b). It also excites a Rossby wave with dispersion to the northeast direction along the South Asian wave guide, which induces cyclonic circulation and cold temperature around the northern Indian subcontinent and western TP, providing a favorable condition for occurrence of snowfall and deepening of snow depth over the CWTP (Figure 13b). Actually, heating over the WIO alone could also excite the northeastward propagating Rossby wave (Figures 13b and 13d). The atmospheric circulation and temperature response to cooling over the WMC are weak over the TP (figure not shown).

The atmospheric circulation and temperature to the east of 90°E show similar responses to the cooling over the WNP with those to the combined heating over the three regions (Figures 13a and 13c). They are also similar to the patterns of circulation and temperature anomalies associated with the Niño-3 index (e.g., cyclonic circulation and cold temperature from the ETP to the WNP) (Figures 10e, 13a, and 13c).

The above simulation results suggest that convection anomalies over the WNP (WIO) works as an important medium by which ENSO (IOD) affects the ETP (CWTP) snow depth. Nevertheless, convection anomalies over the WMC do not exhibit direct impact on snow depth variability over the CWTP.

5. Conclusions

While some studies reported that ENSO exerts a significant role in snow anomalies over the TP (Shaman & Tziperman, 2005; Wang & Xu, 2018), other studies argued that the snow variability over the TP is modulated by IOD rather than by ENSO (Yuan et al., 2009, 2012; Zhang et al., 2019). In this study, we are trying to answer three questions: (1) Can ENSO and IOD affect snow depth over the TP? (2) If yes, what are the
independent roles of ENSO and IOD on snow depth anomalies over the TP? (3) How do ENSO and IOD affect snow depth over the TP?

Climatological snow depth starts to build up in September at high altitudes over the TP, continues to increase in the subsequent months, peaks in January, and decays thereafter. Decrease in surface air temperature and increase in snowfall contribute to deepening of snow before January. However, the decrease in snow depth after January is mostly caused by the increase in surface air temperature.

On interannual time scale, there is significant correlation of ENSO/IOD with snow depth over the TP in early winter (November–December). This is different from the results of obtained by Yuan et al. (2009, 2012) and Zhang et al. (2019), which demonstrated a negligible influence of ENSO on the early winter TP snow cover/snow depth. The discrepancies might be due to the different data sets and variables used in the studies. Further, it is found that ENSO (IOD) has an independent influence on early winter snow depth over the ETP (CWTP).

In early winter, the interannual variability of snow depth over the TP is highly correlated with surface air temperature and snowfall, with a higher correlation with the latter. ENSO is accompanied by an anomalous cyclonic circulation and an anomalous cold center to the southeast of TP in the middle and upper troposphere. The anomalous circulation transports extra moisture air to the ETP, together with the cold anomalies over the ETP, favoring above-normal rainfall (snowfall) and deepening of snow depth over the ETP. These circulation and temperature anomalies can be directly excited by diabatic cooling of suppressed convection anomalies over the WNP as a Rossby wave response. It is well known that ENSO suppresses convection over the WNP (Wang et al., 2000). Thus, ENSO favors deepening of snow depth over the ETP by suppressing convection over the WNP (Wang et al., 2000).

On the other hand, convection anomalies over the WIO associated with IOD could generate a barotropic Rossby wave that propagates northeastward along the South Asian wave guide, consistent with Yuan et al. (2009, 2012). This Rossby wave induces an anomalous cyclonic circulation across the northern CWTP, which transports more moisture to the CWTP from the tropics, providing a favorable condition for occurrence of snowfall and deepening of snow depth over the region. The convection anomalies over the WMC associated with IOD, however, do not exhibit direct impact on snow depth variability over the CWTP.

References


