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RESEARCH ARTICLE

Possible contribution of the inter-annual Tibetan Plateau snow cover variation to the Madden–Julian oscillation convection variability

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The Madden–Julian oscillation (MJO) is the major mode of intra-seasonal variability in the Tropics, yet researches about factors related to this oscillation have not been well understood, particularly on an inter-annual timescale. As a crucial index of the thermal condition of the plateau, the Tibetan Plateau snow cover (TPSC) has attracted great attention with its potential climatic effects. In this paper, observational evidences present that the inter-annual variation in the wintertime TPSC can exert influence on the convection variability associated with the MJO. For the excessive TPSC winters, the MJO convection tends to be stronger in the Indian Ocean (phases 2–3), while it is more vigorous over the western Pacific (phases 6–7) during the reduced TPSC winters. Furthermore, the South Asia High (SAH) is found to be the key system in the potential physical mechanism. A reduced (excessive) TPSC can excite upper-level anomalous anticyclone (cyclone) over the Tibetan Plateau and eastern China, which is in favour of westwards extension (eastwards withdrawal) and enhancing (weakening) of the SAH over tropical oceans. Through an anomalous tropical zonal-vertical circulation, the intensified ascending (descending) motions over the tropical western Pacific may subsequently depress (enhances) the convection over the Indian Ocean. Over the Indian Ocean (phases 2–3) and the western Pacific (phases 6–7), the zonal-vertical circulation associated with the MJO is exactly superposed with the inter-annual anomalous vertical circulation linked with the TPSC, so the according changes of MJO convection are most significant in these two categorized phases.

KEYWORDS

inter-annual variability, Madden–Julian oscillation, South Asia High, Tibetan Plateau snow cover anomaly

1 | INTRODUCTION

The Madden–Julian oscillation (MJO) is known as one of the most significant modes of the intra-seasonal variability in the Tropics (Madden and Julian, 1971). Coupled with deep convection, it propagates eastwards along the equator with a period of 30–60 days.

As an important signal in the Tropics, the MJO has direct influence on the tropical weather and climate, for example, the onset, active, and break of tropical monsoons (Krishnamurti and Subrahmanyam, 1982; Tong *et al.*, 2008), the genesis of tropical cyclone (Wang and Moon, 2017), the variation of surface wind over ocean (Foltz and McPhaden, 2004), and the multi-scale changes of precipitation in the Tropics (Benedict and Randall, 2007; Vincent *et al.*, 2016). Moreover, it can also modulate the weather and climate systems in the extratropical regions. Researches show that the mid-latitude atmospheric circulations may receive anomalous angular momentum from the MJO in the Tropics (Anderson and Rosen, 1983). This intra-seasonal variability induces anomalous horizontal circulation, vertical motion, and moisture supply over East Asia and thus causes anomalous wintertime air temperature and precipitation (Jeong *et al.,* 2005; 2008). The MJO can also initiate polewards propagating Rossby waves and then influence the main modes of low-frequency variability in mid-and-high latitudes, for example, the North Atlantic oscillation (NAO), the Arctic oscillation (AO), and the Pacific North American (PNA) teleconnection pattern (Higgins and Mo, 1997; Cassou, 2008; L'Heureux and Higgins, 2008; Johnson and Feldstein, 2010; Seo and Son, 2012). It is worth mentioning that there might be a two-way interaction between the MJO and mid-latitude oscillation. Lin *et al*. (2009) suggested that the NAO variability may result in the tropical upper zonal wind and then affect the initiation of the MJO in the Indian Ocean.

Although the MJO is renowned as an intra-seasonal variation, it also exhibits prominent year-to-year variability in its behaviour (Weickmann, 1991; Hendon *et al.,* 1999; Pohl and Matthews, 2007). It appears to be very strong and coherent in some years but weak in others. Particularly in Pacific, zonal wind of the MJO presents prominent inter-annual variability in the lower than upper troposphere (Gutzler, 1991). It is actually a challenging topic to reveal what controls the inter-annul variability in the intensity, the number of events, as well as active region of the MJO. Several forcing such as vertical wind shear or diabatic heating, have been put forwards to explore the possible mechanism of the MJO behaviour (Lau and Chan, 1988; Lim *et al.,* 1991; Bladé and Hartmann, 1993). The major driver is sea surface temperature (SST), including El Niño-Southern Oscillation (ENSO) and Indian Ocean dipole (IOD). As the most remarkable inter-annual signal in the Tropics, ENSO has been suggested that it might modulate the inter-annual variation of the MJO. Fink and Speth (1997) pointed out that the MJO activity over the Pacific near equator is compactly related to the SST anomalies and the phase of ENSO. Their results were supported by Slingo *et al*. (1999), whose study showed that the MJO activity is more vibrant during the cold phase of ENSO, but it would also be more energetic with the tropical SST increasing due to global warming. Besides, Hendon *et al*. (1999) detected an eastwards shift of MJO activity east of the date line during El Niño events. However, they claimed that the overall level of MJO activity has nothing to do with El Niño generally, but still feebly link to SST anomalies in the equatorial Indian Ocean and western Pacific. In addition, during the negative phase of IOD, the MJO activity is enhanced with strong convection and organized wind anomalies over the Indian Ocean and Maritime Continent, but subdued during positive IOD (Wilson *et al.,* 2013). These differences are especially distinct over the eastern Indian Ocean, where IOD has the greatest impact on low-level humidity and modulates the MJO afterwards. Izumo *et al*. (2010), who found two types of MJO with different frequency based on satellite observations, revealed that these MJO events can also be modulated by IOD through the background atmospheric circulation or ocean–atmosphere coupling over the southwest Indian Ocean. Therefore, the factors that can affect the inter-annual variation of the MJO, along with its underlying mechanisms are still less well documented.

The Tibetan Plateau, the highest plateau in the world, is called as the "Roof of the world" or the "The Third Pole," and it is viewed as a huge heat/cold source of the global atmosphere (e.g., Luo and Yanai, 1983; Yanai and Li, 1994; Yeh and Wu, 1998). Thermodynamic activities of the Tibetan Plateau change the global atmospheric circulations and have significant influence on the Asian monsoon system (Wu and Zhang, 1998; Zhang *et al.,* 2004), East Asian precipitation (Hsu and Liu, 2003; Wang *et al.,* 2008), and ENSO (Wu *et al.,* 2012). Snow cover can prominently change the thermal condition of the atmospheric underlying surface and thus affect the weather and climate around (Hahn and Shukla, 1976; Goodrich, 1982; Barnett *et al.,* 1989; Groisman *et al.,* 1994; Chen *et al.,* 2000; Qian *et al.,* 2003; Wu and Qian, 2003). As a crucial measure of the thermal condition of the plateau, the Tibetan Plateau snow cover (TPSC) is more important than other areas with its miraculous altitude reaching the middle troposphere (Lin and Wu, 2012). The strong inter-annual variability in the TPSC leads to anomalies of atmospheric diabatic heating at the middle troposphere (Lin and Wu, 2011) and, as a result, affects the regional, even the global climate. For instance, the sensible heating effects of the TP in winter generate an asymmetric dipole zonal-deviation circulation, with a large anticyclone gyre to the north and a cyclonic gyre to the south (Wu *et al.,* 2007). It then enhances the cold outbreaks from the north over East Asia, resulting in a dry climate in South Asia and a moist climate over the Indochina Peninsula and south China. In winter 2008–2009, the severe drought event in northern China was suggested to have connection with the heat forcing of TP (Gao and Yang, 2009). The temperature above normal over the TP, which lessens the moisture transportation from the Bay of Bengal to eastern China, combined with La Niña contributed to this extreme weather. Liu *et al*. (2009) found that Asian monsoon changes during the late Holocene are forced by changes in both solar output and oceanic–atmospheric circulation patterns, whose mechanisms operate not only in low latitudes but also in the northern TP. Moreover, the wither TPSC even can exert impacts on the Asian monsoon (Zhang and Tao, 2001) and ENSO teleconnection (Wu *et al.,* 2012) in the next summer by changing the surface heating, and even can explain Eurasian heat wave with 30% variances (Wu *et al.,* 2015).

The Tibetan Plateau is lying to the north of the Indian Ocean, where the MJO initiates and the convection related to the MJO is strongest during its cycle. It is reasonable to speculate that the TPSC might influence the MJO convection activity. In addition, a recently published report (Li *et al.,* 2016) reveals the influence of the MJO on the TPSC on an intra-seasonal timescale. Their results show that the TPSC increase/decrease with the MJO suppressed/enhanced convection located over the Maritime Continent, but there is non-significant change of TPSC in other phases. As the MJO–NAO relationship, is there any two-way interaction between the MJO and the TPSC? What is the inter-annual reflection of this relationship? Can the TPSC modulate the MJO activities? And what kind of physical mechanism is responsible for such modulation? In this paper, we will focus on the TPSC–MJO relationship in the boreal winter, in which the MJO is most prominent.

This paper is organized as follows. In section 2, we describe our data sets and methods used in this study. The observational TPSC anomaly evolution in wintertime from 1979 to 2012 and the relationship between the TPSC and the MJO are presented in section 3. Section 4 investigates the modulation of the wintertime TPSC on the MJO convection on an inter-annual timescale and the underlying mechanism in detail. Finally, we summarize our findings in the last section.

2 | DATA AND METHODOLOGY

The real-time multivariate MJO (RMM) index was developed by Wheeler and Hendon (2004) to monitor and forecast the MJO in operation, which is available from the Australia Meteorology Bureau (see online at http://www.bom.gov.au/climate/mjo/). Although it might not be the best index of the MJO, this index is convenient to study comparability between the MJO and other variables and this fact has been verified among various works indeed. Defined by the RMM index, an MJO cycle is divided into eight phases, and to be more concise, categorized into four phase groups. Phases 2–3 denote that the MJO convection is centred over the Indian Ocean; phases 4–5, the Maritime continent; phases 6–7, the western Pacific; and phases 8–1, the Western Hemisphere and Africa. Generally, the MJO is featured by a large-scale zonal-vertical cell propagating eastwards with the maximum convective activity that occurs over the warm waters of the Indian Ocean and western Pacific, whereas the MJO is less well coupled with convection in the Western Hemisphere (Zhang, 2005). Figure 1 is an example diagram of the RMM index for December 1, 2007 to February 31, 2008. The daily RMM1 and RMM2 indexes define the state of the MJO with a point in a two-dimensional phase space, and the anticlockwise trace in sequential days indicates the eastwards propagation of the MJO. This figure also shows the relative intensity of the MJO quantified by $\sqrt{\mathrm{RMM1}^2+\mathrm{RMM2}^2}$. The weak MJO with its amplitude less than 1 is located inside the

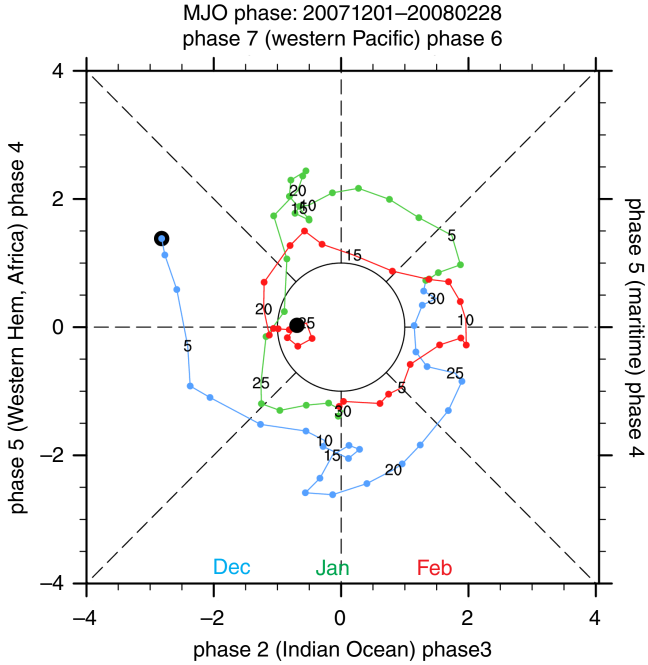

FIGURE 1 Two-dimensional phase space defined by RMM1 and RMM2 (Wheeler and Hendon, 2004), in which the number of co-ordinates implies the value of RMM index. The polygonal line, which indicates the observed MJO index in the winter from December 1, 2007 to February 28, 2008, presents the eastwards propagating of the MJO event [Colour figure can be viewed at wileyonlinelibrary.com]

circle. In this study, composite analyses are all based on the anomalous strong MJO cycles.

Monthly snow cover data set obtained from the Global Snow Laboratory (Rutgers University) has a matrix with 89 × 89 grids of polar stereographic projection of the Northern Hemisphere in each month from 1979 to 2012. Other data sets used in this study are (a) monthly and daily ERA-Interim reanalysis data, including meridional wind, zonal wind and geopotential height; (b) monthly and daily outgoing long-wave radiation (OLR) data obtained from the National Centers for Environment Prediction-National Center for Atmospheric Research (NCEP-NCAR). Both of the data sets have a global coverage with a resolution of 2.5 × 2.5° spanning from 1979 to 2012. The boreal wintertime is defined as from December 1 in last year to February 28 in the present year. To get the components associated with the MJO, we applied 30–60-day band-pass filter on daily data by using Lanczos filtering (Duchon, 1979).

3 | RELATIONSHIP BETWEEN THE TPSC AND THE MJO

Figure 2a presents the climatology of the TPSC in wintertime (DJF) from 1979 to 2012. There is a large snow cover centre over the eastern TP, similar to the results of other researches (e.g., Lin and Wu, 2011). Following Lin and Wu (2011), this area will be used to quantitatively measure the year-to-year variation of the wintertime TPSC. So, we define a snow cover index TPSCI as the snow cover anomaly averaged within the eastern TP domain (90°–105°E, 25°–50°N) shown by a rectangle box in Figure 2a. The time series of the normalized TPSCI for the past 34 winters (1979–2012) illustrates a strong year-to-year variability in the TPSC, along with a pronounced increasing trend (Figure 2b). As a strong signal of inter-annual variability in the Tropics, ENSO may exert profound impacts on the global climate (e.g., Gershunov and Barnett, 1998; Alexander *et al.,* 2002; Wu and Lin, 2012; Wu and Zhang, 2015, among others), thus may simultaneously affect both the MJO and the TPSC. To get the inter-annual variability of the TPSC without the influences of the long-term trend and ENSO, the increasing trend and the ENSO impact with Niño3.4 index are linearly removed from the original TPSCI (dashed in Figure 2b) and this TPSCI is applied in the following study.

In the present work, the standard deviation of ±0.5 is used to distinguish the above/below-normal TPSC winters. Accordingly, nine excessive snow cover winters (1980, 1982, 1985, 1992, 1999, 2002, 2004, 2007, 2010), and 11 reduced snow cover winters (1979, 1981, 1990, 1993, 1994, 1995, 1996, 2000, 2005, 2008, 2012) are selected. Moreover, the RMM indices are used to categorize the phases and amplitudes of the MJO during the excessive snow cover winters and reduced snow cover winters, respectively. The behaviours of the MJO in the anomalous TPSC winters will then be explored by statistic and composite.

As mentioned above, a MJO cycle could be divided into eight phases, which correspond to vigorous convection over four sectors. For simplicity, the contiguous phases are composited as phase groups (i.e., phases 2–3, phases 4–5, phases 6–7, and phases 8–1). Table 1 shows the number of days and averaged amplitude for categorized phase groups in the excessive, reduced TPSC winters and in long-term climatology, respectively. The MJO tends to prefer remaining from Indian Ocean to Pacific (phases 2–7), with more vitality in the long run. During the excessive TPSC winters, the strong MJO is more likely to be found in phases 6–7 with less occurrences in phases 8–1. The MJO intensity measured by the averaged RMM indices is strongest in

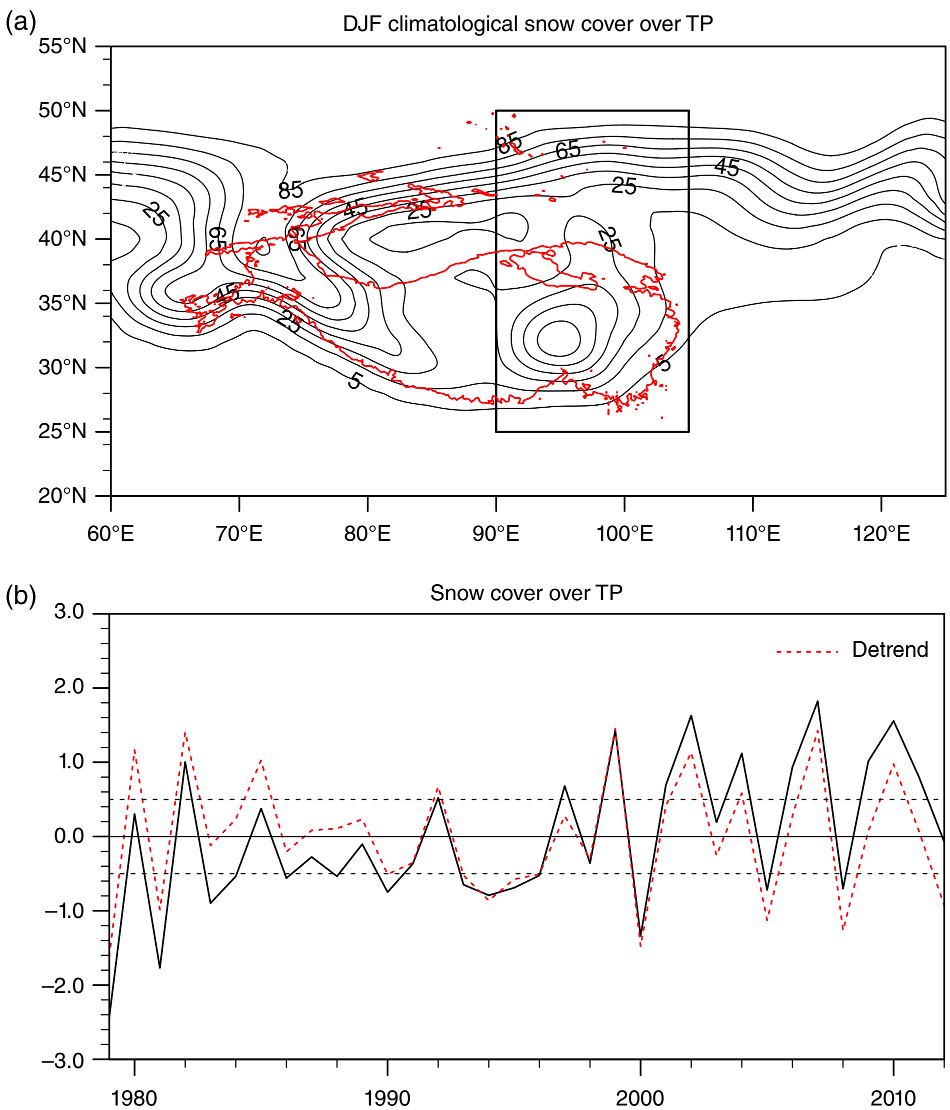

FIGURE 2 (a) Wintertime (December–February, DJF) mean snow cover (%) over the TP for the period of 1979–2012. The areas included in the curves are 2,500 m above sea level. (b) The normalized TPSCI time series (solid curve) defined by the winter TPSC anomaly averaged in the boxed area (90°–105°E, 25°–50°N) in (a) from 1979 to 2012. The dashed indicates the detrended component of the TPSCI and the ENSO impact is also removed by linearly detrending Niño3.4 index. The dashed horizontal lines denote ±0.5 standard deviation of the normalized TPSCI [Colour figure can be viewed at wileyonlinelibrary.com]

TABLE 1 Number of days and averaged amplitude for categorized phases of the MJO in the excessive, reduced TPSC winters and in long-term climatology, respectively

	Phase	2–3	4–5	6–7	8–1
Excessive TPSC winters	No. of days	91	109	172	73
	Mean amplitude	1.992	1.581	1.684	1.639
Reduced TPSC winters	No. of days	135	202	178	99
	Mean amplitude	1.550	1.704	1.800	1.563
In long-term climatology	No. of days	495	483	599	381
	Mean amplitude	1.746	1.700	1.712	1.672

phases 2–3, and weakest in phases 4–5. On the contrary, during the reduced TPSC winters, most number of days of the MJO occurs in phases 4–5, while the averaged MJO intensity is strongest in phases 6–7. It is worthy to be mentioned, compared to that in the reduced TPSC winters, the MJO intensity in the excessive TPSC winters is much stronger (weaker) when it passes by the Western Hemisphere and the Indian Ocean (Maritime Continent and the western Pacific), that is, phases 8–3 (4–7). Especially, the difference of mean amplitude between these two cases in phases 2–3 is up to 0.44, which is greatest among all the categorized phases. That means, the MJO convection is much stronger for the excessive TPSC winters when it passes through the Indian Ocean, where is to the south of the Tibetan Plateau. Does it imply the effect of the TPSC on the MJO?

To obtain a more specific understanding of the difference in MJO between the typical excessive and reduced TPSC winters, the composite distributions of OLR in each phase group are shown in Figure 3. Both in the excessive and reduced TPSC winters, the MJO convection features eastwards propagation along the equator from the Indian Ocean to the Pacific Ocean. However, the convection anomalies over the Indian Ocean exhibit great differences. The positive anomaly of convection is much stronger in the excessive TPSC winters when the MJO active convection is located over the Indian Ocean in phases 2–3. In phases 6–7, the suppressive convection associated with the MJO is located over the Indian Ocean and the negative anomalies of convection is more obvious in the reduced TPSC winters, as well as the intensity of MJO shown in Table 1. The differences of OLR over the Indian Ocean and western Pacific domain between high and low TPSCI winters may illustrate such features more clearly (Figure 4). In phases 2–3 and phases 6–7, there are negative OLR centres over the tropical Indian Ocean. In other phases, the differences in OLR are not very obvious.

Here come the questions, why is the MJO convection over the Indian Ocean during the excessive TPSC winters stronger than that in the reduced TPSC winters? How does the TPSC modulate this intra-seasonal oscillation on an inter-annual timescale during wintertime? We will try to give an interpretation in next section.

4 | POSSIBLE PHYSICAL MECHANISMS

As the highest plateau in the world, the Tibetan Plateau plays an important role in the global weather and climate variation due to its thermal effect. The changes in the snow cover may lead to significant variation of the thermal status of the plateau. It is notable that air temperature anomalies at 2 m (shading in Figure 5) in the excessive TPSC winters are almost contrary to those in the reduced ones. Over the eastern Tibetan Plateau, surface temperature is lower than normal in the excessive TPSC winters. At the same time, snow cover anomalies (contour in Figure 5) during the excessive and reduced TPSC years centre on eastern TP and to its north, which accord with the distribution of temperature anomaly. These features further certify that the region we selected as key region to measure TPSCI is reasonable, and the heating effect of snow cower is significant.

Li *et al*. (2016) suggested that there is maximum TPSC in phases 2–3 but with minimum anomalies in these phases on the intra-seasonal timescale. Therefore, it is reasonable that the TPSC in phases 2–3 is stable on the intra-seasonal timescale and we can eliminate the probability that the variability of MJO strength in these phases can affect TPSC. In this article, the strength of MJO in phases 2–3 is significant different during the excessive and reduced TPSC winters, which infers that TPSC can exert impact on the MJO. So we conclude that the subtropical convection activities play a passive role, rather than active one to the middle-latitude TPSC anomalies in phases 2–3. To explore the TPSC-related atmospheric circulation anomalies at difference levels, the zonal-vertical sections of geopotential height anomalies along 30°N for the excessive and reduced TPSC winters are calculated (Figure 6). For the excessive TPSC winters, when the TP is a relative cold source to the atmosphere, deep negative anomaly of geopotential height can be found over the main part of TP and to its east with a centre located at 200 hPa. Thus, the relative cold surface over the eastern TP in the excessive TPSC winter cools the atmosphere down, leading to the subsidence of isobaric surface (Figure 6a). Oppositely, in the reduced TPSC winters, the relative higher surface temperature is in favour of the positive geopotential height anomaly (Figure 6b). And the difference of geopotential height anomaly (Figure 6c) reveals significant otherness over the eastern TP, corresponding to the temperature anomaly at 2 m.

As presented in Figure 6, the anomalous centres of geopotential height are located at the upper level around 200 hPa. Consequently, the anomalous geopotential height and winds at 200-hPa for the excessive and reduced TPSC winters are shown in Figure 7. In the excessive TPSC

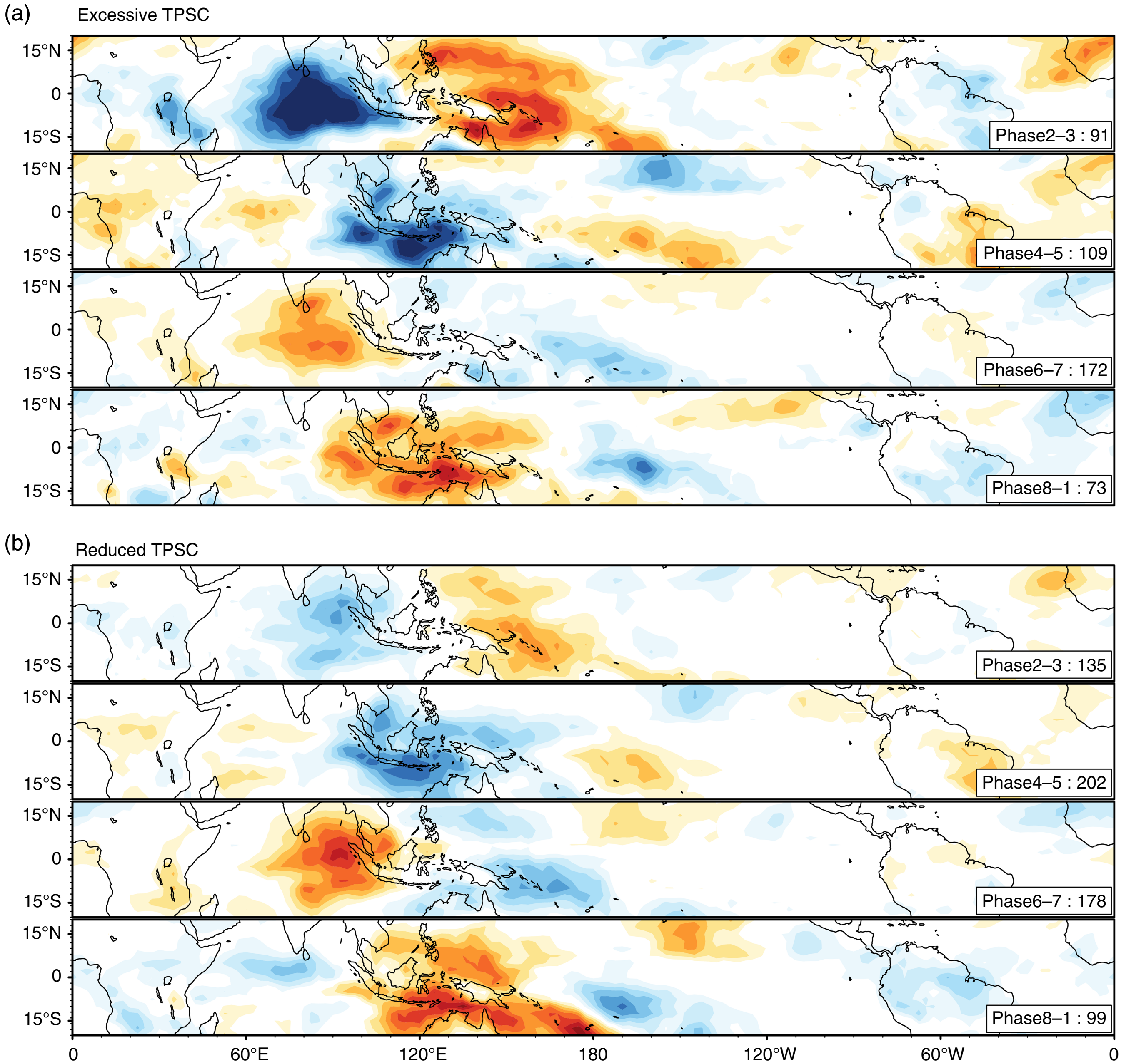

FIGURE 3 Evolution of the composite OLR (W/m^2) for the excessive (a) and reduced (b) TPSC winters [Colour figure can be viewed at wileyonlinelibrary.com]

winters (Figure 7a), a negative geopotential height band is seen over the subtropical region with significant anomaly centre taking up eastern TP and eastern China, corresponding with the key region of anomalous temperature and TPSC. Correspondingly, an anomalous cyclone appears over the eastern TP and regions to its east with distinct northerly anomalies occupying the eastern TP. Meanwhile, we can see a strong anomalous anticyclone centre coupled with positive geopotential height anomalies in the high latitudes north of the TP. For the reduced TPSC winters, opposite atmospheric circulation patterns are presented with an anomalous anticyclone to the east of TP and the southerly prevailing over the eastern TP (Figure 7b). Difference of these two cases (Figure 7c) shows the same pattern as the former, especially the puissant northwesterly associated with the cyclone over eastern Asia are blowing across the equator from north to south.

It is not difficult to find that those TPSC-induced anomalous atmospheric circulations primarily cover the tropical to subtropical regions of eastern Asia (Figure 7), where is mainly controlled by the South Asia High (SAH) over the tropical oceans in winter (Qian *et al.,* 2002; Huang and Qian, 2007). The SAH is a bond connecting the tropical and subtropical systems at high levels, and might be regarded as a potential linkage between the TPSC and the MJO. Figure 8 calculates the composites of emblematic line of the SAH identified as 12,200 gpm for the excessive and reduced TPSC winters. It is indicated that the SAH might

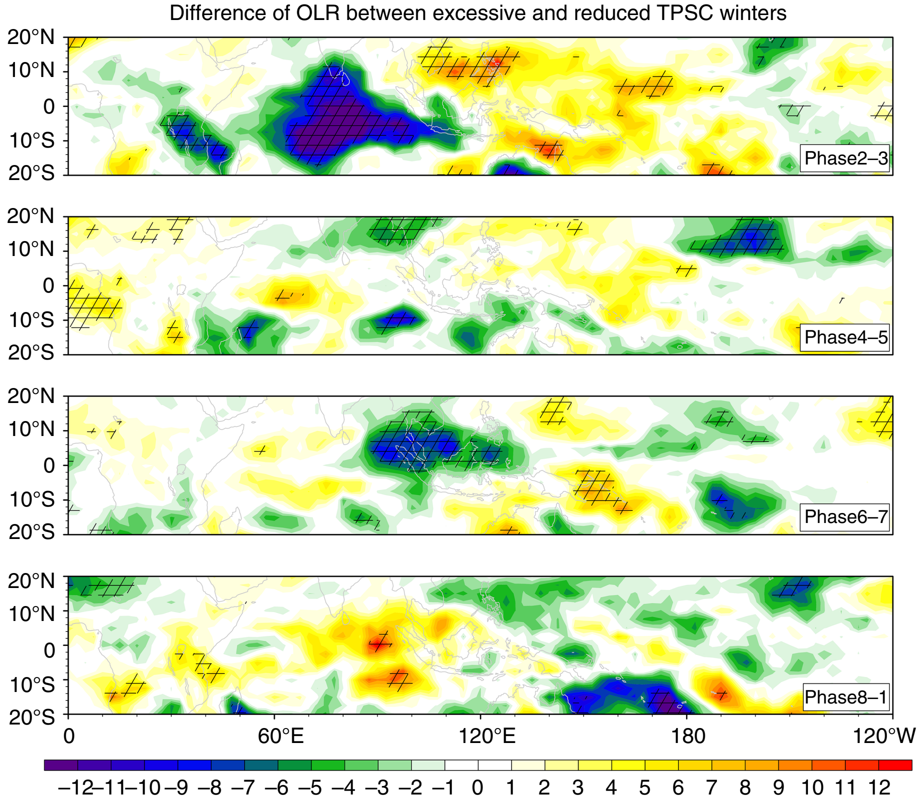

FIGURE 4 Difference in OLR between the excessive and reduced TPSC winters. The dashed area with slash indicates 99% confidence level based on the Student's t test [Colour figure can be viewed at wileyonlinelibrary.com]

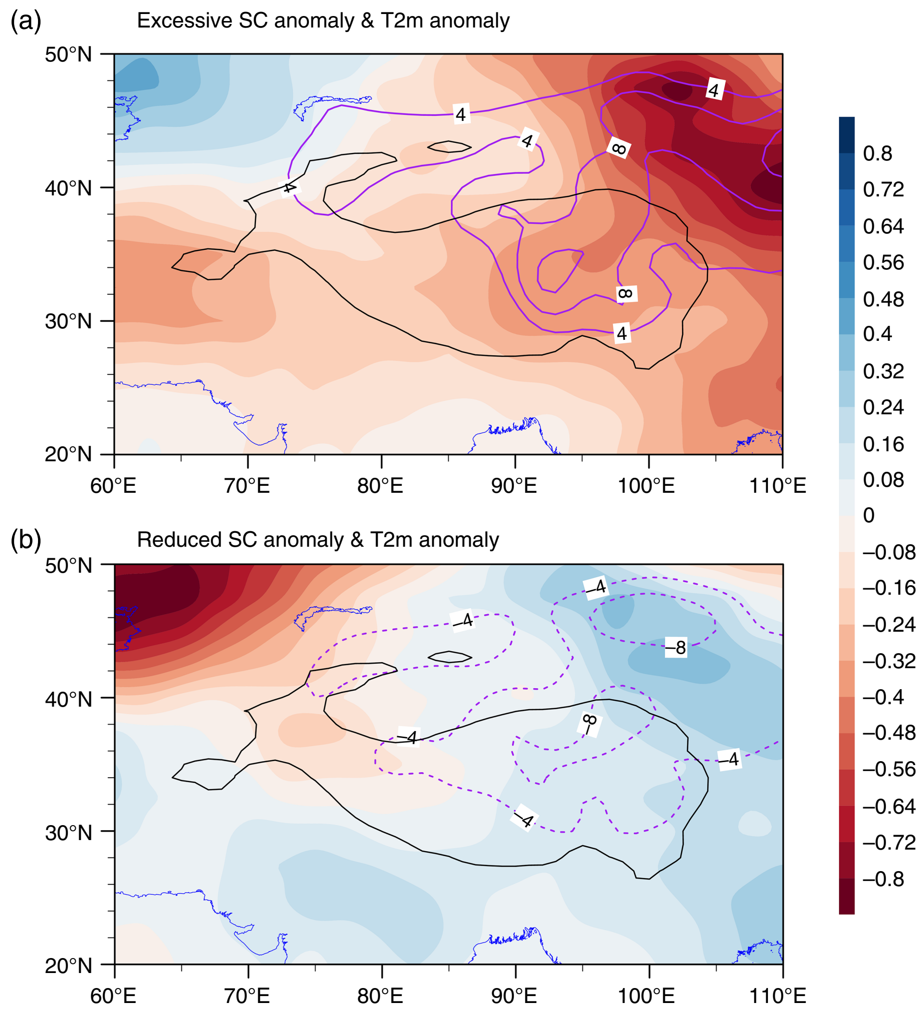

FIGURE 5 DJF temperature anomalies at 2-m (shading in units of°C) and anomalous snow cover (contour in units of %) for the (a) excessive and (b) reduced TPSC winters in the TP domain [Colour figure can be viewed at wileyonlinelibrary.com]

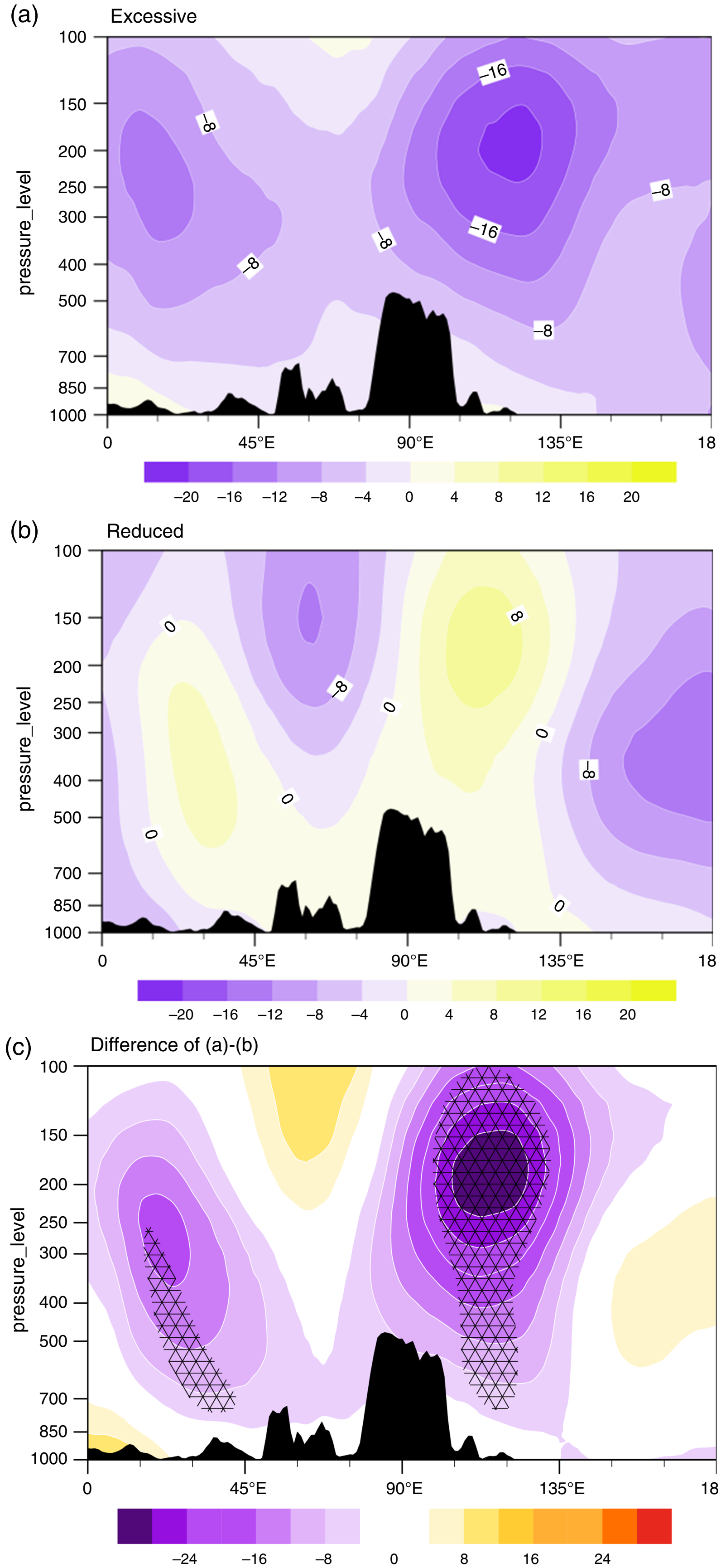

FIGURE 6 DJF zonal circulation of geopotential height anomalies (shading in units of gpm) along 30°N for the (a) excessive and (b) reduced TPSC winters and the difference (c) in the TP domain. The contour interval is 4 gpm. The dash area with slash in (c) indicates 95% confidence level based on the Student's *t* test. The black shading in middle and lower troposphere indicates towering mountains, which can reveal topograph of TP succinctly and clearly [Colour figure can be viewed at wileyonlinelibrary.com]

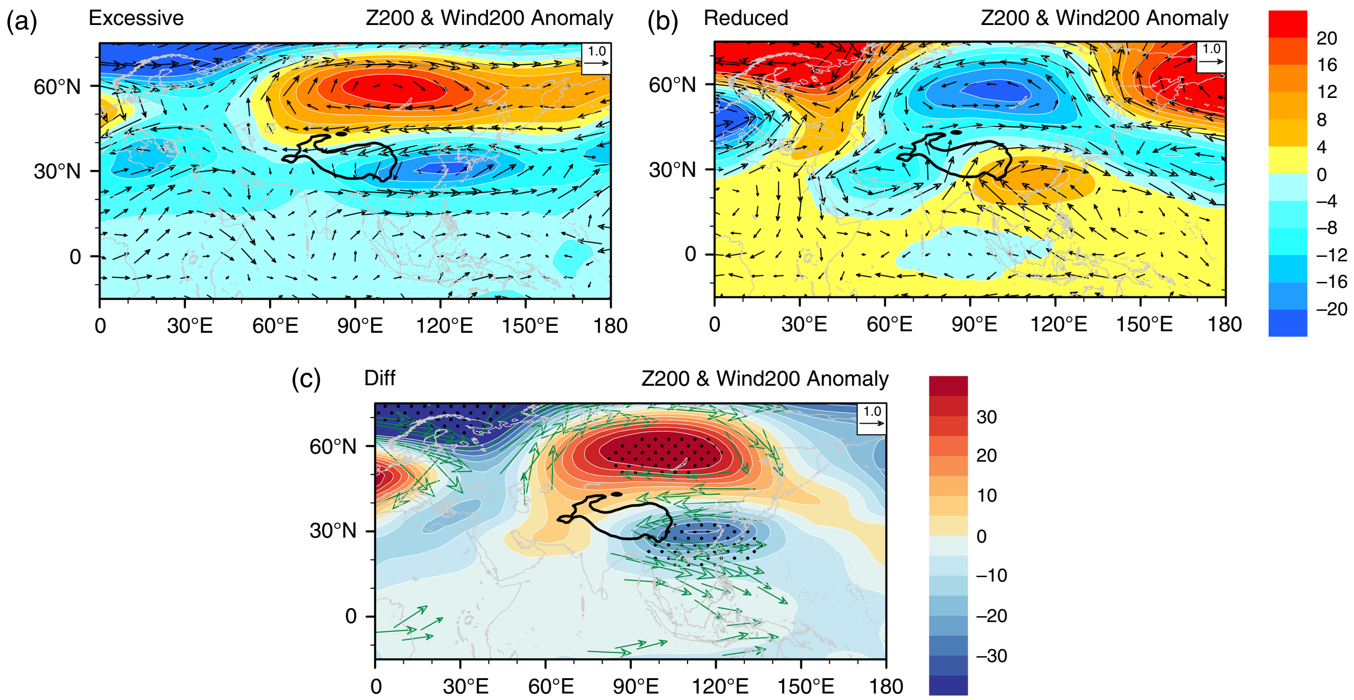

FIGURE 7 DJF geopotential height anomalies (shading in units of gpm) and winds (vectors in units of m/s) anomalies at 200 hPa for the (a) excessive, and (b) reduced TPSC winters. The contour interval is 4 gpm. The dotted areas and vectors in (c) exceed the 95% confidence level based on the Student's *t* test [Colour figure can be viewed at wileyonlinelibrary.com]

shift farther westwards in the reduced TPSC winters than in the excessive TPSC cases, which is in agreement with the corresponding positive circulation anomalies at 200 hPa (Figure 7b).

Coincided with the movement of the SAH, velocity potential anomalies and divergent wind anomalies at 200-hPa associated with the TPSC are investigated (Figure 9), as well as the vertical motion and convection in the tropical region (Figure 10). In the excessive TPSC winter, it is the convergence over the Tropics that might suppress the convection and ascending motion, although the signals are not as remarkable as those in the reduced winters (Figure 9a). For the reduced TPSC winters, notable anomalous divergence over the tropical western Pacific at high levels is favourable for the enhancement of ascending motion and convection (Figure 9b). Besides, the difference (Figure 9c) presents the same phase from the eastern TP to the western Pacific with significant winds converging over the Tropics.

Further, zonal-vertical sections of vertical circulation along the Tropics (averaged over 10°S–10°N) (Figure 10) displays anomalous descending/ascending flows over the western Pacific Ocean coincided with the convergence/divergence in Figure 9. It is indicated that the changes in vertical circulation over the Pacific Ocean would induce the circulation over the Tropics to its west. During the excessive TPSC winters, the ascending anomalies near 90°E can be found with the descending anomalies in the east, while opposite conditions are found in reduced TPSC winter. When comes to the difference, a powerful and conspicuous zonal cell can be detected with ascending anomalies near 90°E and the descending anomalies in the east. It is obvious that these anomalous vertical motions would be favourable

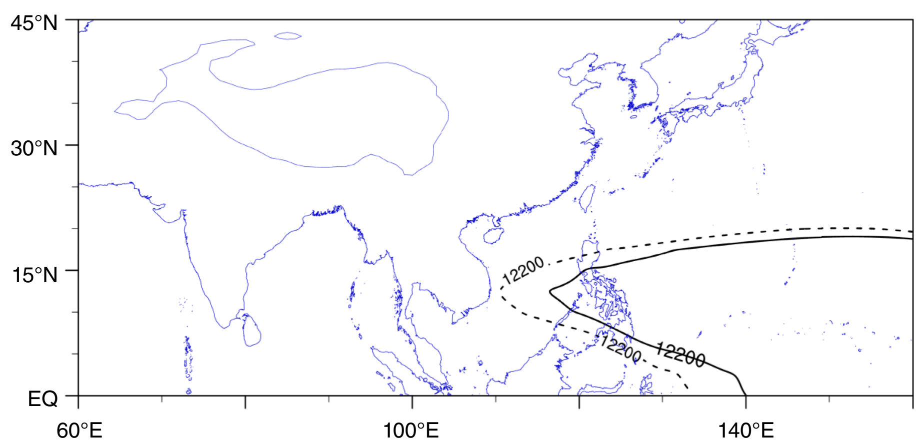

FIGURE 8 Composite 12,200 gpm contour at 200 hPa for the excessive (solid line) and reduced (dotted line) TPSC winters, which can represent the feature line of the SAH in winter [Colour figure can be viewed at wileyonlinelibrary.com]

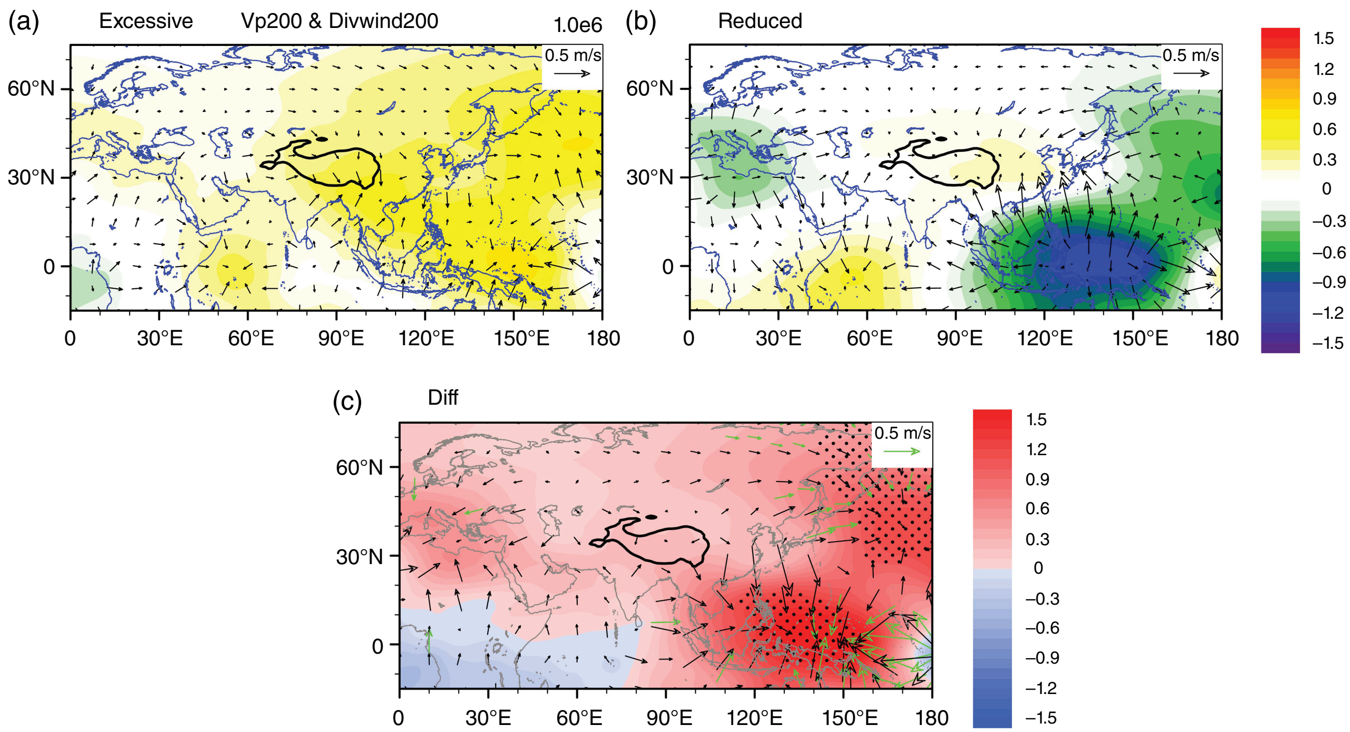

FIGURE 9 DJF velocity potential anomalies (shading) and divergent wind anomalies (vectors in units of m/s) at 200 hPa for the (a) excessive and (b) reduced TPSC winters. The dotted areas and light-colored vectors in (c) exceed the 95% confidence level based on the Student's *t* test [Colour figure can be viewed at wileyonlinelibrary.com]

(unfavourable) for the enhancement of background vertical motion. Although convection is the source to motivate MJO, can the background vertical motion also contribute to the development of the MJO? In order to investigate the modulation role of the background vertical motion on the development of MJO, following Zhang *et al*. (2015), we calculate the power spectra of the OLR anomalies averaged in the Indian Ocean (60°–100°E, 10°S–10°N) in the excessive and reduced TPSC years, respectively (Figure 11). The power spectrum in the excessive TPSC winters (Figure 11a) shows active MJO activity at the period of 30–45 days, which also exceeds the 95% confidence level statistically. On the contrary, there is no prominent MJO signal during the reduced TPSC winters. It is noticeable that the suppressed convection over the Indian Ocean in the reduced TPSC winter is unfavourable for the occurrence of MJO and the MJO activity is weak, while the enhanced convection in the excessive TPSC winter is accompanied with a significantly active MJO activity. The conclusions above are also can be found in the study of Zhang *et al*. (2015).

To certify this hypothesis, composites of zonal circulation of the MJO along the Tropics (averaged over 10°S–10°N) by categorized phases are examined for the excessive and reduced TPSC winters (Figure 12). In phases 2–3, the composite ascending motion over Indian Ocean in the excessive TPSC winters is much stronger than that in the reduced TPSC winters. On the contrary, in phases 6–7, the composite descending motion over Indian Ocean in the excessive TPSC winter is feebler than that in the reduced TPSC winter. The changes of vertical motion shown here are consistent with the changes of convection shown in Figure 3. In other phases of the MJO (not shown), the centres of active convection (updraft) or suppressive convection (downdraft) are not located over the Indian Ocean. That is, they could not be superposed with the background flows, so that no significant change is found in anomalous TPSC winters.

In brief, the TPSC anomaly can drive the SAH in the upper troposphere over the Pacific by cooling/heating the atmospheric column and then leading to the changes in the upper-level geopotential height and circulation over the TP. During the reduced TPSC winters, the anomalous anticyclonic circulations over eastern Asia may lead to the strengthening and westwards extending of the SAH over the tropical oceans, which is beneficial to the divergence over the western Pacific. With the strong background ascending motion anomalies, the development of convection associated with the MJO cycle over the Indian Ocean is depressed through the transmission of zonal cell, and vice versa.

5 | CONCLUSION AND DISCUSSION

Located to the north of the Indian Ocean, the TP is an enormous cold (or heat) source in Asia with significant influence on the global climate (e.g., Luo and Yanai, 1983; Yanai and Li, 1994; Yeh and Wu, 1998). Also, the MJO initiates from the tropical Indian Ocean and the MJO convection is strongest here during its cycle. Hence, there might be a potential

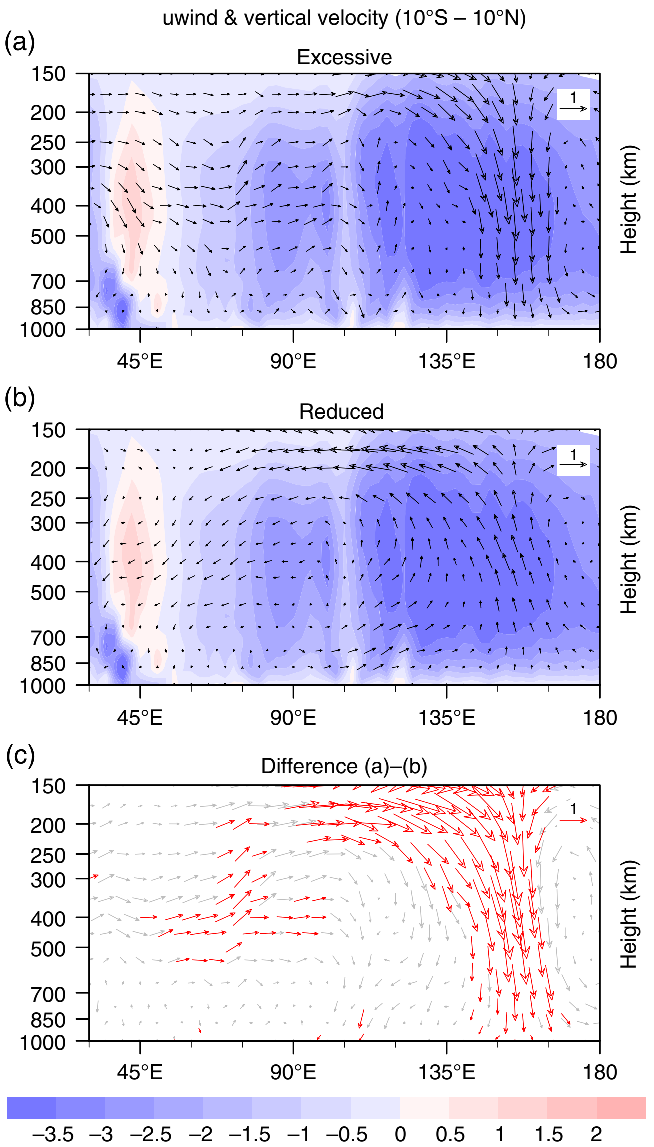

FIGURE 10 Composite DJF zonal circulations anomaly along the Tropics (averaged over 10°S–10°N) for the (a) excessive, (b) reduced TPSC winters and their differences. The vectors include the vertical and zonal components and the dark ones are that exceed the 90% confidence level based on the Student's *t* test. The shaded area is the climatological vertical motion during 1979–2012 winters [Colour figure can be viewed at wileyonlinelibrary.com]

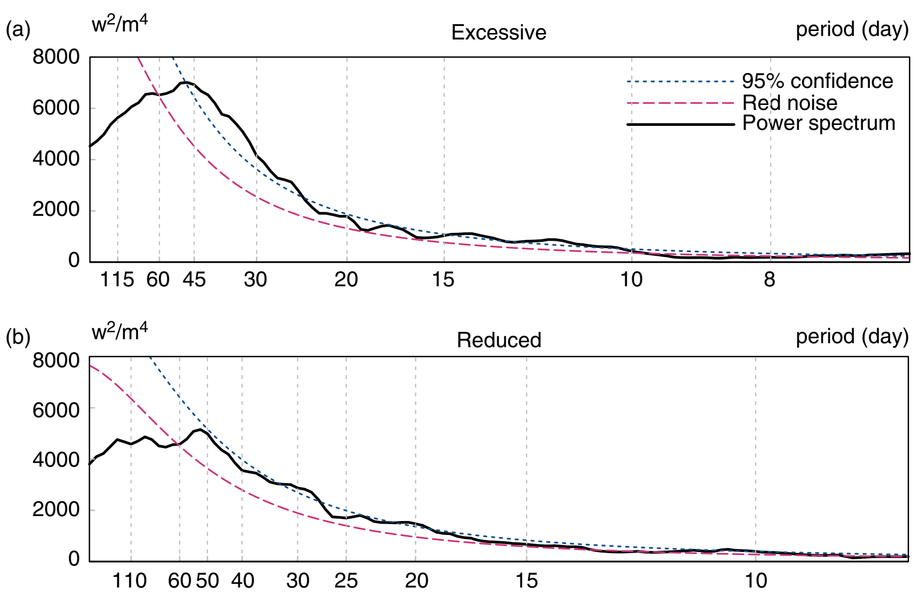

FIGURE 11 Power spectra of the OLR anomalies averaged over 60°–110°E, 15°S–15°N in the (a) excessive and (b) reduced TPSC winters. The thick black and dotted lines represent power spectrum, 95% confidence level and red noise, respectively [Colour figure can be viewed at wileyonlinelibrary.com]

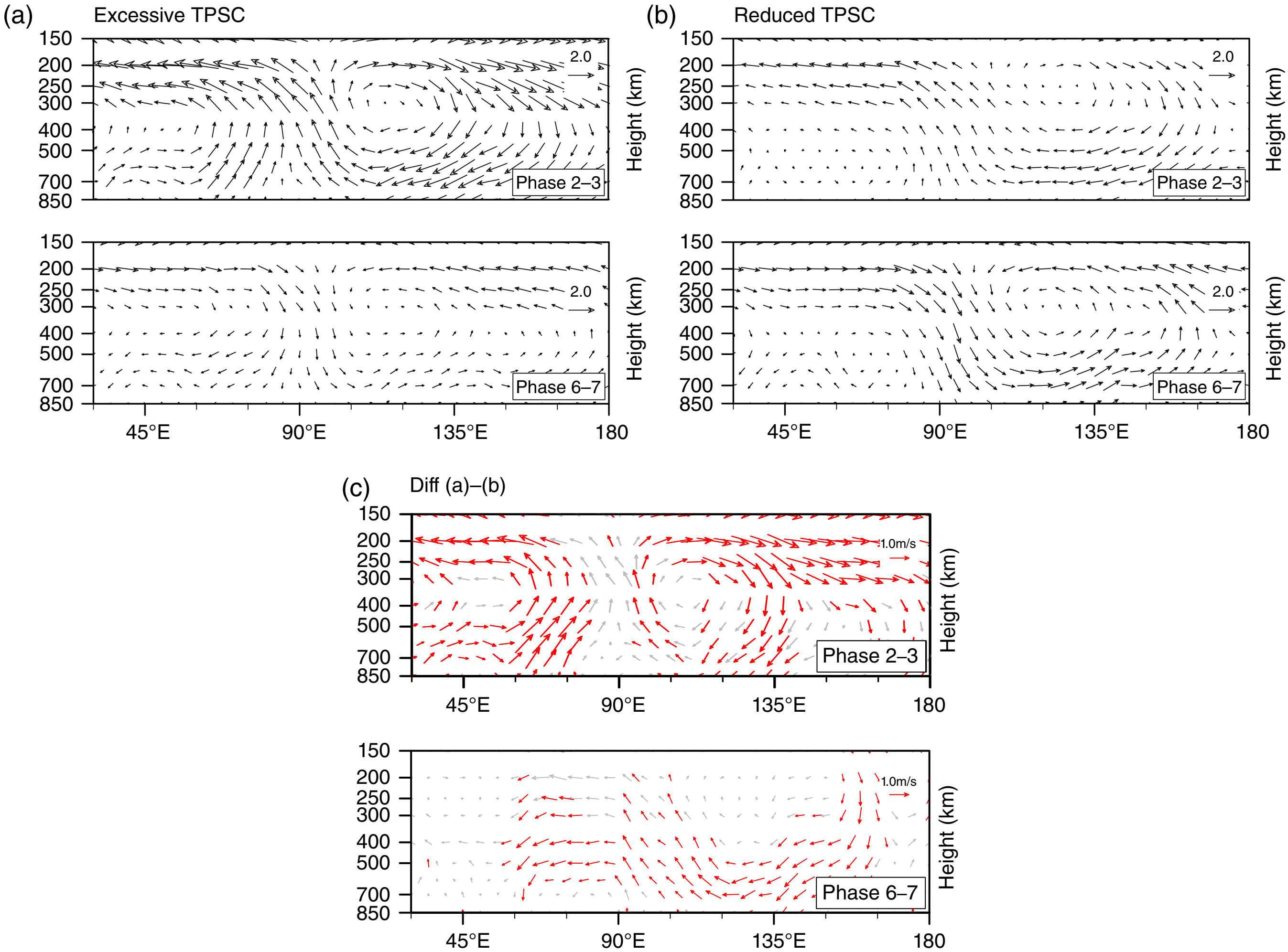

FIGURE 12 Composite DJF zonal circulations associated with the MJO along the Tropics (averaged over 10°S–10°N) by categorized phases for the (a) excessive and (b) reduced TPSC winters. The third one (c) indicates the difference between (a) and (b). The dark vectors exceed the 99% confidence level based on the Student's *t* test [Colour figure can be viewed at wileyonlinelibrary.com]

connection between the TPSC and the MJO in wintertime. Li *et al.* (2016) found that there is indeed a link between the MJO and the wintertime TPSC on an intra-seasonal timescale. In this study, we explored the possible influence of the TPSC on the development of MJO convection in wintertime.

The snow cover over the eastern TP and its adjacent region has prominent inter-annual variability in winter. The excessive (reduced) TPSC winters are chosen based on the normalized TPSCI greater (less) than 0.5 (−0.5) standard deviation, and finally, there are 9 (11) samples can meet the requirement. Besides, two contiguous out of all the eight MJO phases are categorized into four groups and their differences are analysed. Results show that the MJO convection has significant differences between the excessive and reduced TPSC winters in phases 2–3 and phases 6–7. More specifically, in the excessive (reduced) TPSC winters, the MJO convection will be stronger (weaker) than normal when it is located over the Indian Ocean.

Possible physical mechanisms are also investigated. The TP is a huge heat (cold) source for atmosphere in the reduced (excessive) TPSC winters, and can lead to the anomalous anticyclone (cyclone) over the eastern TP at 200 hPa. The SAH, a puissant anticyclonic system that located over the Pacific Ocean in winter, is affected by this anomalous TPSC-induced circulation. Composite analysis shows that the SAH can extend westwards in the reduced TPSC winters and withdraw eastwards in the excessive TPSC winters. For the reduced (excessive) TPSC winters, these changes of the SAH contribute to the divergence (convergence) over the western tropic Pacific, as well as the ascending (descending) motion in this area. And then the convection over the Indian Ocean is exactly depressed (stimulated) through the zonal cell, and so is the MJO convection. In phases 2–3 and phases 6–7, the zonal-vertical circulation associated with the MJO is exactly superposed with the inter-annual anomalous vertical circulation anomalies linked with the TPSC, making the MJO convection most significant in these two phases.

This work suggests that the TPSC can influence the MJO convection on an inter-annual timescale. As a primary mode of intra-seasonal variability in the Tropics, the MJO also has notable inter-annual variabilities that might affect the thermal condition of the TP in return. Therefore, if we take bidirectional relationship into consideration, how does the MJO exert influence on the TPSC on an inter-annual timescale? Besides, we only mentioned circulation anomalies over the eastern TP in Figure 7, but there is significant difference in high latitude between the excessive and reduced TPSC winter. The further mechanism of the circulation anomalies in north of TP are not investigated. Does this anomalous circulation also result from the heating effect of the TPSC? Or will the system in higher latitude also contribute to this high since the anomalous wind in north of the high is derived from ocean and higher-latitude? These issues are worthwhile more exploration and deeper understanding.

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