Based on the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-Interim reanalysis data from 1979 to 2016, an index for Tibetan Plateau (TP) soil moisture ($I_{TPSM}$), which could represent inter-annual variation in spring soil moisture, is defined after removing the linear trend in soil moisture. The late spring soil moisture variation over the TP and its influences on the plateau summer monsoon are analysed. The results show that the TP soil moisture variation is non-uniform, and spring soil moisture over the northeastern TP shows an increasing trend, while the western TP (especially the peripheral area) renders a significantly decreasing trend. When the $I_{TPSM}$ shows positive (negative) anomalies (i.e., spring soil moisture over the northeastern TP is higher (lower), soil moisture over the western TP is lower (higher) than normal), the plateau summer monsoon may be stronger (weaker). Anomalous variations in spring TP soil moisture can eventually affect the plateau summer monsoon by changing atmospheric circulation, diabatic heating, and water vapour transportation. It is found that when the $I_{TPSM}$ exhibits positive anomalies, diabatic heating over the centre and eastern TP is remarkably enhanced, and warm and humid moisture from the Bay of Bengal and Arabian Sea and cold air from high latitudes merge over the central and eastern TP; coupled with lower-level convergence and upper-level divergence, all these contribute to an increase in plateau summer monsoonal precipitation. When the $I_{TPSM}$ shows negative anomalies, diabatic heating over the centre and eastern TP is weakened, the anomalous cyclone to the south of the TP hinders warm and humid air flows transportation to the TP, and the anomalous anticyclone to the southeast of Lake Baikal weakens cold air to the south; coupled with weakened convergence and upwards motion, these factors all lead to a decrease in plateau summer monsoonal precipitation.

**KEYWORDS**
mechanism, plateau monsoon, soil moisture, Tibetan Plateau

1 INTRODUCTION

The Tibetan Plateau (TP), with a mean altitude that exceeds 4,000 m and reaches one third the height of the tropopause, is the highest and most extensive geographic features in the world (Ge et al., 2017). The dynamic and thermal forcing caused by the TP has a great influence on regional and global climate (Ye and Gao, 1979; Wang et al., 2016). In the free atmosphere, the TP is a special thermodynamic system; its seasonal variation in thermal effects can lead to a seasonal pressure system. Shifts in cold and heat sources during winter and summer over the TP can also cause seasonal transformation of the cold high and the warm low; wind fields over the TP that are suitable for pressure fields can also occur seasonal evolutions, which result in a unique plateau monsoon phenomenon (Xu and Gao, 1962; Ye and Gao, 1979). The plateau monsoon occurs in the middle of the troposphere, and its maximum thickness in summer is
2–3 km away from the ground, which plays an important role in weather and climate over the TP even though it is only a phenomenon in the planetary boundary layer. It can function as a bridge between the monsoon in the lower troposphere and the planetary wind system in the upper troposphere, and it also destroys the planetary pressure and wind belt by reinforcing the monsoon in the lower troposphere; these result in various synoptic and climatic characteristics of the TP. The activity range and intensity of plateau monsoon can determine the location of the rain band and the monsoonal rainfall amount, which has a great effect on droughts and floods in the plateau monsoon region and its adjacent areas (Ye and Gao, 1979; Tang, 1995). Overall, the plateau monsoon has a decisive influence on the weather and climate in the TP and its neighbouring areas.

Soil moisture is a key factor for characterizing land surface conditions and it also plays an important role in land–atmosphere interactions. By changing the underlying surface parameters, such as thermal capacity and albedo, soil moisture can influence surface sensible heat and latent heat transport, which can eventually affect local climate change (Ma et al., 2001; Yang et al., 2016). As a slow variable with long-term “memory,” soil moisture anomalies will inevitably have a lasting impact on climate by changing the hydrothermal exchange between the land and atmosphere (Zuo, 2007). The soil moisture “memory” refers to the timescale that the soil is able to maintain an abnormally wet or dry state after the abnormal external forcing has disappeared; it is an important indicator of the land–atmosphere interactions and has crucial significance for seasonal forecasting. Changes in soil moisture can affect not only atmospheric stratification stability but also large-scale atmospheric circulation (Shang and Wu, 2015). Wang and Shang (2005) found that the soil hydrothermal variation not only has an important effect on changes in atmospheric circulation and the Asian summer monsoon but also is an important factor for seasonal precipitation. Considering its importance in water and energy exchanges between land surfaces and the atmosphere, soil moisture is recognized as an essential climate variable (Qiu et al., 2016; GCOS Secretariat, 2010), and the knowledge of its temporal trends and variability is of essential significance for understanding the effects of climate change over the TP.

Prior studies have mainly focused on the contributions of soil moisture anomalies to Asian or African monsoon, rather than to plateau monsoon. For instance, it is found that soil moisture variability can significantly affect the African monsoon system (Douville, 2002; Berg et al., 2017). The pre-monsoonal soil moisture has a remarkable influence on the Indian summer monsoon (Shakeel et al., 2012). Also, there exists significant sensitivity of East Asian summer monsoon to wet soil from the lower and middle reaches of the Yangtze River to north China (Zuo and Zhang, 2016). Zhang and Zou (2011) used both observed and reanalysis data to show that spring soil moisture has a significant impact on the summer monsoon circulation over East Asia. Moreover, the East Asian early winter monsoon is shown to be significantly correlated with soil moisture in the previous summer in eastern China (Liu et al., 2012). The plateau summer monsoon has received much less attention than the South and East Asian summer monsoon even though variations in the three systems are interlinked over various timescales (Duan et al., 2011). Studies about the influence of spring soil moisture on the latter plateau summer monsoon have rarely been discussed. Our previous study indicated that late spring TP soil moisture is significantly correlated with the plateau summer monsoon; when the spring soil moisture over the central and eastern TP is higher (lower) and soil moisture over the western TP is lower (higher) than normal, the plateau summer monsoon may be stronger (weaker) (Zhou et al., 2016); a statistical correlation between the summer plateau monsoon index and preceding spring local soil moisture was simply discussed. However, the possible mechanism that the late spring TP soil moisture affecting the plateau summer monsoon are still unclear.

In this paper, our analyses are focused on the late spring soil moisture variation over the TP and its influences on the plateau summer monsoon, with the aim to explore possible clues how the spring soil moisture affect the plateau summer monsoon and to account for soil moisture variability. The paper is organized as follows: data and methodology are described in section 2. The variation of soil moisture is shown in section 3. Also in this section, we present the possible mechanism of late spring soil moisture anomalies influencing on the plateau summer monsoon and some further discussions. Finally, the conclusions are provided in section 4.

2 | DATA AND METHODOLOGY

2.1 | Data

The data used in the study are derived from the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Reanalysis (ERA-Interim) and have a resolution of 0.5 × 0.5°. The data are from 1979 to 2016 and include monthly mean volumetric water content, geopotential height, specific humidity, surface sensible and latent heat flux, surface temperature, air temperature, 10 m wind speed, and U and V winds. Soil moisture data are classified into four layers: 7, 21, 72, and 189 cm.

With respect to the reliability of the reanalysis data, some scholars (Qi et al., 2009; Hua et al., 2012; Zhou et al., 2016) have used ECMWF reanalysis data to calculate plateau monsoon indices. Therefore, it is feasible to utilize ERA-Interim data for the study of plateau monsoons. Ground-based observation sites in western China, especially over the TP, are scarce, the observation start times are late, and numerous values are missing; these factors all lead to the low availability of data. Furthermore, the timescale of the soil moisture observation data used in the study as of
now is approximately 10 years. Consequently, for regional-scale or long-term climate research, model-derived soil moisture data are often utilized as an alternative (Zuo and Zhang, 2008; Zhou et al., 2016). Previous evaluations of several reanalysis data sets showed that the ERA-40 reanalysis could generally capture the inter-annual variations of soil moisture from the surface to deeper layers over China (Li et al., 2004; Zhang et al., 2008). Reasonable performances for ECMWF soil moisture analyses in the humid monsoon period were found using in situ measurements over the TP (Su et al., 2013; Zeng et al., 2016). Liu et al. (2015) indicated that the ERA-Interim soil moisture has a better performance over the TP compared to the National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR). The ERA-Interim data set used in this study is the latest global ECMWF reanalysis data and adopts the same land surface model as ERA-40, the Tiled ECMWF Scheme for Surface Exchanges over Land (TESSEL). Moreover, the data assimilation system of ERA-Interim has been upgraded from that of ERA-40 to a four-dimensional variational analysis (4D-Var). Model parameters have been modified, and more satellite and observational data have been employed to correct and improve the data accuracy (Dee et al., 2011).

Our research mainly focuses on inter-annual variability of soil moisture over the TP, so it is better to use a relatively longer term in situ measurement to verify the ERA-Interim soil moisture. However, the relatively longer timescale of continuous measured soil moisture over the TP can be obtained is 10 years. In this study, measured soil moisture from the Maqu network, which located at the northeastern edge of the TP (33°30′–34°15′N, 101°38′–102°45′E) and at the first major meander of the Yellow River, are utilized to be compared with that of ERA-Interim. The network, consisting of 20 stations in an area of approximately 40 by 80 km, monitors continuously the soil moisture at different depths (from 5 to 80 cm below surface) at 15 min intervals (Su et al., 2013). Detailed information for the Maqu network can be found in Su et al. (2013) and Chen et al. (2016). Observed data from nine stations of the network at 5 and 10 cm depths (from 5 to 80 cm below surface) at 15 min intervals are selected to validate the soil moisture product from ERA-Interim during the period 2008–2017. The averaged soil moisture for Maqu network and ERA-Interim in late spring (April–May) and summer (June–August) are plotted in Figure 1. Overall, the ERA-Interim reanalysis overestimates the surface and subsurface soil moisture both in spring and summer. Su et al. (2013) indicated that the overestimation in ECMWF results may be attributed to the land surface analysis system used (including the boundary conditions, initialization, soil properties, and the soil physics process and so on) and the soil texture that determines the soil hydraulic properties at Maqu network. The high correlations (Cor, see Figure 1, lower right corner) between the observations and ERA-Interim, which significantly exceed the 95% confidence level, suggest that the inter-annual variations of the in situ measurement can be captured by the soil moisture analysis.

2.2 Methodology

2.2.1 Wavelet analysis

The wavelet analysis is a mathematical analysis method that has been developed in recent years. Compared with the traditional Fourier analysis method, wavelet analysis can decompose time series into the form of a time–frequency field, which provides more information for a localized section of the time series.

For the wavelet function \( \psi(t) \in L^2(R) \), where \( \int_{-\infty}^{+\infty} \psi(t)dt=0 \), the continuous wavelet format is defined as

\[
\psi_{a,b}(t) = |a|^{-1/2} \psi \left( \frac{t-b}{a} \right),
\]

where \( a, b \in R, a \neq 0 \), and the continuous wavelet transform of the signal \( f(t) \in L^2(R) \) is

\[
Wf(a,b)=|a|^{-1/2} \int_R f(t)\overline{\psi} \left( \frac{t-b}{a} \right) dt,
\]

where \( Wf(a,b) \) is the wavelet transform coefficient, \( \psi(t) \) is the mother wavelet, \( \overline{\psi} \) is the conjugate operator, \( a \) is the scale parameter, and \( b \) is the translation parameter.

In practical application, time series for meteorological elements are generally discrete. For the discrete series \( f(k\Delta t) \) \( (k = 1, 2, \ldots, N) \), where \( \Delta t \) is the sample time interval, Equation (2) can be rewritten as follows:

\[
Wf(a,b)=|a|^{-1/2} \Delta t \sum_{k=1}^{N} f(k\Delta t)\overline{\psi} \left( \frac{k\Delta t-b}{a} \right).
\]

The wavelet variance can be calculated as

\[
Var(a)=\int_{-\infty}^{+\infty} |Wf(a,b)|^2 db.
\]

The wavelet variance can reflect the signal fluctuation energy distribution that changes with scale parameter \( a \). The main periodicities that exist in the time series can be determined by the wavelet variance (Li et al., 2013; Kuang et al., 2014).

2.2.2 Plateau monsoon index

The plateau monsoon index (PMI) is calculated by the following:

\[
\text{PMI} = \text{std} \left[ U'_{(28°–31°N,85°–95°E)} - U'_{(34°–37°N,85°–95°E)} \right] + \text{std} \left[ V'_{(30°–35°N,92.5°–102.5°E)} - V'_{(30°–35°N,77.5°–87.5°E)} \right]
\]

where \( \text{std} \left[ U'_{(28°–31°N,85°–95°E)} - U'_{(34°–37°N,85°–95°E)} \right] \) is the standardized value of the difference of the mean zonal wind
anomalies between \(28^\circ - 31^\circ N, 85^\circ - 95^\circ E\) and \(34^\circ - 37^\circ N, 85^\circ - 95^\circ E\) at 550 hPa, and \(\text{std}[V(30^\circ - 35^\circ N, 92.5^\circ - 102.5^\circ E)] - V(30^\circ - 35^\circ N, 77.5^\circ - 87.5^\circ E)\) is the standardized value of the difference of the mean meridional wind anomalies between \(30^\circ - 35^\circ N, 92.5^\circ - 102.5^\circ E\) and \(30^\circ - 35^\circ N, 77.5^\circ - 87.5^\circ E\) at 550 hPa. A larger PMI value indicates stronger wind shear and cyclonic rotation near the surface over the TP and shows that the plateau summer monsoon is stronger. Meanwhile, a smaller PMI value indicates a weaker cyclonic rotation of the wind field at the surface layer, as well as a stronger plateau winter monsoon (Zhou et al., 2016).

The Mann–Kendall statistical analysis (Smadi and Zghoul, 2006) and the composite analysis (Xun et al., 2012; Ge et al., 2017) are also applied to study the late spring soil moisture variation over the TP and its possible mechanism affecting the plateau summer monsoon.

3 | RESULTS AND DISCUSSION

3.1 | Variation of soil moisture over the TP

The spatial distributions of late spring and summer soil moisture trends over the TP from 1979 to 2016 are shown in Figure 2. The trend distributions for the 7- and 72-cm layers are similar to those for the 21- and 189-cm layers, respectively. Here, we only show soil moisture trend distributions for the shallow 7-cm layer and the deep 189-cm layer. Soil moisture variation over the TP is non-uniform. In late spring, 7-cm layer soil moisture shows decreasing trend in most part of the TP, except in northeastern of the plateau (Figure 2a). The greatest drying regions, located in the Kunlun Mountain and the north of Himalayas, have rates smaller than \(-0.02 \text{ m}^3/\text{m}^3/10\text{a}\), which significantly exceed the 90% confidence level. For the deep layer spring soil moisture (Figure 2c), the decreasing trend over the western TP is weaker, while the increasing trend over the northeastern TP is stronger than that for the shallow layers. In summer, shallower layer soil moisture presents a significant increasing trend over the past 38 years except for some small regions over the northwestern and southeastern TP. The most prominent wetting areas occur over the Chi-lien Mountain, as well as to the south side of the Kunlun Mountain, with rates greater than 0.01 \text{ m}^3/\text{m}^3/10\text{a} that significantly exceed the 90% confidence level. The spatial distribution of deep layer summer soil moisture trend is similar to that of spring, but with some regional differences; this illustrates that abnormally deep layer soil moisture in spring can be maintained until summer which, in turn, could affect the summer thermal forcing and atmospheric circulation over the TP and its adjacent areas. Because the study mainly focuses on the linkages between inter-annual variations in spring soil moisture and the plateau summer monsoon, the linear trends of soil moisture are removed from the following study.

For the soil moisture variability over the TP, diverse conclusions have been drawn based on different soil moisture data sets. Wang et al. (2016) found that the soil moisture over the TP exhibits an increasing trend using satellite retrieval data from 1988 to 2010, indicating that there exist spatial differences in spring soil moisture over the TP and that the soil moisture trends over the eastern and western TP exhibit obvious opposite phase variation characteristics. However, significantly decreasing trend is detected in the soil moisture over the TP using observation soil moisture

3.2 | Differences of late spring and summer soil moisture between the TP and adjacent areas

The greatest drying regions, located in the Kunlun Mountain and the north of Himalayas, have rates smaller than \(-0.02 \text{ m}^3/\text{m}^3/10\text{a}\), which significantly exceed the 90% confidence level. For the deep layer spring soil moisture (Figure 2c), the decreasing trend over the western TP is weaker, while the increasing trend over the northeastern TP is stronger than that for the shallow layers. In summer, shallower layer soil moisture presents a significant increasing trend over the past 38 years except for some small regions over the northwestern and southeastern TP. The most prominent wetting areas occur over the Chi-lien Mountain, as well as to the south side of the Kunlun Mountain, with rates greater than 0.01 \text{ m}^3/\text{m}^3/10\text{a} that significantly exceed the 90% confidence level. The spatial distribution of deep layer summer soil moisture trend is similar to that of spring, but with some regional differences; this illustrates that abnormally deep layer soil moisture in spring can be maintained until summer which, in turn, could affect the summer thermal forcing and atmospheric circulation over the TP and its adjacent areas. Because the study mainly focuses on the linkages between inter-annual variations in spring soil moisture and the plateau summer monsoon, the linear trends of soil moisture are removed from the following study.

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from agro-meteorological stations during the period 1981–2010 (Zhang et al., 2016). As stated above, there is no consensus reached on the TP soil moisture temporal–spatial variations yet. Considering the reasonable performance of ERA-Interim soil moisture data (Su et al., 2013; Liu et al., 2015; Zeng et al., 2016), the results of TP soil moisture variability in this study are based on the ERA-Interim reanalysis.

From the above analysis, we can see that the spatial distribution in spring soil moisture trends from 1979 to 2016 is non-uniform. The large standard deviation areas, which coincide with the large inter-annual variability regions for the TP spring soil moisture, mostly lie in the northeastern, southeastern, and western TP (Figure 2e). Therefore, areas with significant rising trends, as shown in Figure 2a (35°–38°N, 92°–102°E), are selected to calculate the regional mean soil moisture in April and May. After normalization, the soil moisture index over the northeastern TP ($I_{NESM}$) is obtained. The soil moisture index over the western TP ($I_{WSM}$) can also be obtained using the same method, but in a significantly decreasing region (30°–36°N, 75°–85°E). The time series for $I_{NESM}$ and $I_{WSM}$ are displayed in Figure 3a. It can be seen that both time series present fluctuating inter-annual variation from 1979 to 2016. To quantitatively represent the inter-annual variation in spring soil moisture over the TP, $I_{NESM}$ minus $I_{WSM}$ is simply defined as the comprehensive plateau soil moisture index ($I_{TPSM}$). The inter-annual variation in $I_{TPSM}$ and the plateau summer monsoon index (PSMI) from 1979 to 2016 are given in Figure 3b. Overall, the $I_{TPSM}$ and PSMI exhibit similar inter-annual and inter-decadal variation characteristics. The $I_{TPSM}$ is significantly and positively correlated with the PSMI. The inter-annual variation in the $I_{TPSM}$ is in alignment with that of the PSMI, with a correlation coefficient of 0.38 that significantly exceed the confidence level at 95%. This demonstrates that when the $I_{TPSM}$ exhibits positive (negative) anomalies (i.e., spring soil moisture over the northeastern TP is higher (lower), soil moisture over the western TP is lower (higher) than normal), the plateau summer monsoon may be stronger (weaker), which is consistent with results provided in the previous study (Zhou et al., 2016).

To further analyse the late spring soil moisture variation over the TP, the wavelet analysis is carried out to the $I_{TPSM}$. When the real part of the wavelet transform coefficient is positive (negative), it indicates that soil over the northeastern TP is suffering from a wet (dry) duration, while the western

![FIGURE 2](https://wileyonlinelibrary.com)
TP is suffering from a dry (wet) duration; the coefficients are plotted with solid (dotted) lines in the figure. The wavelet real part shows the peak and valley structures for the $I_{TPSM}$ on different timescales. The real part of the Morlet wavelet transformation is shown in Figure 3c. It is found that the phase structure of soil moisture changes on different timescales and presents alternatively positive or negative variations over time. In addition, it suggests that there are quasi-periodic variations on the following timescales in the $I_{TPSM}$ series: 3–8, 9–16, and 16–26 years. To confirm the main periods that determine the variation of the $I_{TPSM}$ series, variance of the wavelet transform coefficient is calculated, as shown in Figure 3d. There are three peak values, which are located along the timescale at 6, 13, and 22 years, illustrating that the $I_{TPSM}$ series has three periodic oscillations of 6, 13, and 22 years. The largest peak value corresponds to the timescale at 22 years, illustrating that the time period of 22 years fluctuates the most; therefore, it is considered as the first main period. In addition, peak values at 13 and 6 years are considered the second and third main periods, respectively.

### 3.2 The possible mechanism of spring soil moisture over the TP affecting the plateau summer monsoon

Years from 1979 to 2016 with $I_{TPSM} \geq 1$ ($I_{TPSM} \leq -1$) are selected as positive (negative) anomaly years; then, 7 years (1980, 1981, 2004, 2006, 2007, 2008, and 2009) are characterized as positive anomaly years, while the other 6 years (1989, 1994, 1995, 1997, 2001, and 2003) are characterized as negative anomaly years. Therefore, the possible mechanism that the spring soil moisture over the TP affecting the plateau summer monsoon can be discussed using composite atmospheric circulation, surface sensible and latent flux, and integrated vapour flux. Note that the long-term linear trends in the meteorological data have been removed.

The composite height and wind field anomalies at 850, 500, and 200 hPa are exhibited in Figure 4; anti-phase distributions of anomalous wind and height fields are evident between positive and negative anomaly $I_{TPSM}$ years. During positive anomaly $I_{TPSM}$ years, an anomalous anticyclone is located to the south of the TP at 850 hPa (Figure 4a), and the anomalous wind to the southeast of the TP is southerly, which is favourable for warm and humid water vapour from the Bay of Bengal and Arabian Sea to be transported along the southern edge of the TP to the TP and its surrounding areas. Concurrently, an anomalous anticyclone is located to the southeast of the Lake Baikal, where the anomalous northeasterly wind on its southeast side aids the transportation of cold air from high latitudes to the TP and its adjacent areas. However, during negative anomaly $I_{TPSM}$ years (Figure 4b), an anomalous anticyclone and an anomalous cyclone, in turn, occur from the southwest to the northeast; the anomalous cyclone to the south of the TP hinders the transportation of moisture from Bay of Bengal and Arabian Sea to the TP. Meanwhile, a huge anomalous anticyclone extends from the northeastern side of the TP to the eastern Japan cut-off the cold airflows from the high latitudes. During positive anomaly $I_{TPSM}$ years, the anomalous height field over the TP at 500 hPa is negative and the TP is dominated by an anomalous cyclone (Figure 4c). There exist northeasterly over the northern TP and southwesterly over the southern TP, and the convergence of the wind makes the ascending motion enhanced. An anomalous anticyclone and an anomalous cyclone are located to the northeast of the TP.
and northeastern China, respectively; this cyclone is conducive to the genesis, development, and enhancement of the northeast cold vortex; the anomalous northeasterly between the cyclone and anticyclone accompanying the northeast cold vortex contribute to the transportation of cold air flows from boreal areas to the south, and also enhance anomalous northeasterly over the northern TP remarkably. Meanwhile, an anomalous anticyclone occurs to the south of the TP, which could continuously supply moisture to the TP and strengthen the southwesterly over the southern TP. All these

FIGURE 4 Composites height field (contour, unit: gpm) and horizontal wind field (vectors, unit: m/s) anomalies at 850, 500, and 200 hPa in summer during positive (a, c, e) and negative (b, d, f) TPSM years. The shaded blue (pink) areas indicate the wind (height) fields that have passed the confidence level at 90% [Colour figure can be viewed at wileyonlinelibrary.com]
further enhance warm and humid airflows pushing northwards, which reinforce the uplifting movement and form a circulation condition conducive for monsoonal precipitation over the TP. However, in negative anomaly $I_{TPSM}$ years (Figure 4d), the anomalous height field over the TP at 500 hPa is positive. Under the influence of the anomalous cyclone, which is located to the north and the south of the TP, anomalous southwesterly and northeasterly air flows severally prevail over the northern and southern TP, respectively. There exists a divergence of wind field, which cannot form circulation conditions that contribute to monsoonal rainfall over the TP. In the upper troposphere (200 hPa), a powerful anomalous anticyclone extends from central and eastern TP to the East China Sea for positive anomaly $I_{TPSM}$ years (Figure 4e), making the South Asian High abnormally intensified and eastwards. Concurrently, anomalous cyclones located over the northeastern China and western marginal TP deepen the East Asian trough and the trough to the south of Lake Balkhash, which can reinforce meridional circulation and is conducive to the cold air flows from high latitudes pushing southwards. Opposite to the positive years, an anomalous cyclone dominates over the central and eastern TP (Figure 4f), which, accompanying lower level divergence, is not conducive to the uplifting movement of air flow over the TP, and decreases the plateau summer monsoonal precipitation.

To further analyse the influence of spring soil moisture over the TP on the latter summer atmospheric circulation, the pressure–longitude cross section of the summer vertical velocity and wind of divergence anomalies between 30°N and 36°N for anomaly $I_{TPSM}$ years are shown in Figure 5. For the $p$-velocity anomalies $\omega$, positive (negative) values represent descending (ascending) motion. During positive (negative) anomaly $I_{TPSM}$ years (Figure 5a,b), anomalous strong ascending (descending) motion is detected over the central and eastern plateau with rising (sinking) height approximately up to 200 hPa. For wind divergence anomalies, positive (negative) values represent divergence (convergence) of wind fields. It can be concluded that the wind field over the TP shows a remarkable convergence (divergence) at lower level while divergence (convergence) at upper level for positive (negative) anomaly $I_{TPSM}$ years (Figure 5c,d). It can be deduced that the intensified convergence and anomalous ascending motion result in above-normal monsoonal rainfall over the central and eastern TP for positive anomaly $I_{TPSM}$ years. During negative anomaly $I_{TPSM}$ years, this situation is reversed.

As seen from the above analysis, the anomalous variation in spring soil moisture can influence the latter plateau summer monsoon by changing circulation at upper and lower levels. It is found that when the $I_{TPSM}$ shows positive (negative) anomalies (i.e., the spring soil moisture over the northeastern TP is higher (lower), the soil moisture over the western TP is lower (higher) than normal), the convergence (divergence) in the lower troposphere and the divergence (convergence) in the upper troposphere enhance (weaken) the uplifting movement of the air flows over the central and eastern TP, which forms a circulation condition conducive (not conducive) to the occurrence of plateau summer monsoonal precipitation.

Surface sensible heat and latent heat flux can reflect the hydrothermal exchange between the surface and the atmosphere, as well as have a great influence on atmospheric circulation and regional climate by changing diabatic heating (Hui et al., 2005). The composite surface sensible heat and latent heat flux anomalies in summer are presented in Figure 6. During positive anomaly $I_{TPSM}$ years (Figure 6a), the sensible heat anomalies are positive over the central and eastern TP, and negative over the edge of the western TP. Opposite to the positive years, the distribution of sensible heat anomalies are negative over the eastern TP and positive over the western TP for the negative anomaly $I_{TPSM}$ years (Figure 6b). The possible reason for these abnormal distribution characteristics of the sensible heat anomalies may relate to wind speed and $T_s - T_a$ (differences between surface and air temperature) over the TP. The sensible heat is calculated as $SH = \rho C_p V (T_s - T_a)$, where $C_p$ is the specific heat of dry air at a constant pressure, $\rho$ is air density, $C_d$ is the drag coefficient for heat, $V$ is the wind speed measured at 10 m above the ground, and $T_s$, $T_a$ are the surface and air temperature, respectively (Duan et al., 2011). It can be seen that the sensible heat flux is closely related to the surface wind speed and $T_s - T_a$; the stronger the wind speed and $T_s - T_a$, the larger the flux is. So the composite wind speed (10 m) and $T_s - T_a$ anomalies in summer during positive and negative anomaly $I_{TPSM}$ years are given in Figure 6e–h. During positive anomaly $I_{TPSM}$ years, the central and eastern TP is dominated by an anomalous cyclone, and strong wind convergence leads to the increase in the surface wind velocity; moreover, cloud cover, water vapour, and aerosol, which influence the total energy received by the ground (Shen et al., 2008), may lead to the enhancement of $T_s - T_a$; these factors all result in the abnormal increase of sensible heat flux over the central and eastern TP. However, during negative anomaly $I_{TPSM}$ years, wind velocity, and $T_s - T_a$ decrease relatively, which decreases the surface sensible heat over the central and eastern TP. During positive anomaly $I_{TPSM}$ years (Figure 6c), latent heat anomalies are positive over the eastern and northwestern TP, and negative over the western and northeastern TP, indicating that the atmospheric heat source over the central and eastern TP is mainly dominated by the latent heat of condensation. During negative anomaly $I_{TPSM}$ years (Figure 6d), except for some small areas over the central and northeastern TP, latent heat anomalies over the TP are negative. It can be concluded that the occurrence of monsoonal precipitation over the TP is remarkably increased in positive anomaly $I_{TPSM}$ years compared to that of negative anomaly $I_{TPSM}$ years.
During positive anomaly $I_{TPSM}$ years, diabatic heating over the central and eastern TP is remarkably enhanced, and the eastern TP is controlled by a warm low, which could contribute to the ascending movement of air flows over the central and eastern TP. However, during the negative anomaly $I_{TPSM}$ years, diabatic heating over the central and eastern TP is weakened, which is not conducive for the uplifting movement of the airflows over the central and eastern TP.

Water vapour is one of the prerequisites for monsoonal precipitation. Moisture can be transported to rainfall areas through large-scale air flow movements, and then precipitation can form under certain circulation conditions. Composite integrated vapour flux (from surface pressure to 300 hPa) anomalies in summer are presented in Figure 7. During positive anomaly $I_{TPSM}$ years (Figure 7a), an anomalous anticyclonic vapour transportation circulation extends from south of the TP to the Bay of Bengal; over its western part the anomalous southerly vapour transportation is favourable for conveying large amounts of warm and humid moisture from the Bay of Bengal and Arabian Sea into northern areas. There also exists an anomalous anticyclonic vapour transportation circulation to the southeast of Lake Baikal, which provides a large amount of dry and cold air flows from high latitudes to the TP. Air flows from high latitudes and southern ocean merge over the central and eastern TP, sufficient moisture accompanying lower-level convergence and upper-level divergence, cause an increase in monsoonal precipitation over the central and eastern TP. During negative anomaly $I_{TPSM}$ years (Figure 7b), an anomalous anticyclonic vapour transportation circulation and an anomalous cyclonic vapour transportation circulation in turn occur from southwest to northeast, the relatively weaker anomalous cyclonic vapour transportation circulation hinders the transportation of moisture from southern ocean to the TP. Meanwhile, a huge anomalous anticyclonic vapour transportation circulation hinders the transportation of moisture from southern ocean to the TP. Reduced moisture accompanying weakened convergence and updrafts, could lead to the decline of monsoonal rainfall.

Water vapour and its variations are related to atmospheric activities and climate changes at regional and global scales.
scale, moisture condensation processes at the lower troposphere is essential to latent heat transfer (Held and Soden, 2000; Trenberth and Stepianiak, 2003a; 2003b; Wagner et al., 2006; Gurbuz and Jin, 2017). Atmospheric expansion or compression caused by pressure variations has little effect on the specific humidity, so the specific humidity

FIGURE 6 Composite surface sensible heat (SH) flux, latent heat (LH) flux (unit: W/m²), wind speed (10 m, unit: m/s), and differences between surface and air temperature (T_s – T_a, unit: k) anomalies in summer during positive (a, c, e, g) and negative (b, d, f, h) anomaly TPSM years. The dotted areas have passed the confidence level at 90% [Colour figure can be viewed at wileyonlinelibrary.com]
is often utilized to characterize the water vapour content in the atmosphere (Guo and Ding, 2014). Figure 7c,d shows the pressure–longitude cross section of the specific humidity (unit: $10^{-4}$ kg/kg) anomalies between 30°N and 36°N in summer during (a, c) positive and (b, d) negative anomaly $I_{TPSM}$ years. The shaded and dotted areas have passed the confidence level at 90%. The shaded grey areas in the lower part of figure indicate the terrain height [Colour figure can be viewed at wileyonlinelibrary.com]

From the above analysis, it can be seen that anomalous soil moisture in spring has a remarkable impact on vapour transportation in the summer. When the $I_{TPSM}$ shows positive anomalies, warm and humid water vapour from the Bay of Bengal and Arabian Sea and cold air from high latitudes merge over the central and eastern TP; coupled with lower level convergence and upper level divergence, these could contribute to an increase in plateau summer monsoonal precipitation. When the $I_{TPSM}$ shows negative anomalies, the anomalous cyclonic vapour transportation circulation to the south of the TP hinders warm and humid air flows to the TP, and anomalous anticyclonic vapour transportation circulation to the southeast of Lake Baikal weakens the cold air to the south; coupled with weakened convergence and upwards motion, all these result in a decrease in the plateau summer monsoonal precipitation.

4 | CONCLUSIONS

Based on the ECMWF ERA-Interim reanalysis data from 1979 to 2016, an index for Tibetan Plateau soil moisture ($I_{TPSM}$) that could represent inter-annual variation in spring soil moisture is defined after removing the linear trend in the
soil moisture. The variation of spring soil moisture over the TP and its possible mechanism affecting the plateau summer monsoon are investigated. The conclusions of the study are as follows:

1. The soil moisture variation over the TP is non-uniform; spring soil moisture over the northeastern TP shows an increasing trend, while the western TP (especially peripheral area) renders a significant decreasing trend. In addition, the $I_{TPSM}$ is significantly and positively correlated with the plateau summer monsoon index (PSMI). When the $I_{TPSM}$ exhibits positive (negative) anomalies, that is, the spring soil moisture over the northeastern TP is higher (lower), the soil moisture over the western TP is lower (higher) than normal, the plateau summer monsoon may be stronger (weaker).

2. The phase structures of the spring soil moisture over the TP ($I_{TPSM}$) change over different timescales and present alternatively positive or negative variations over time. There exist quasi-periodic variations along the timescales at 16–26, 9–16, and 3–8 years, and the corresponding major periods are about 22, 13, and 6 years.

3. When the $I_{TPSM}$ exhibits positive (negative) anomalies, convergence (divergence) in the lower troposphere and divergence (convergence) in the upper troposphere enhance (weaken) the uplifting movement of the air flows over the central and eastern TP, forming a circulation condition conducive (not conducive) for the plateau summer monsoon precipitation.

4. During positive anomaly $I_{TPSM}$ years, diabatic heating over the central and eastern TP is remarkably enhanced, and the central and eastern TP is controlled by a warm low, which could contribute to the ascending movement of the air flows over the central and eastern TP. During negative anomaly $I_{TPSM}$ years, diabatic heating over the central and eastern TP is weakened, which is not conducive for the uplifting movement of the air flows over the central and eastern TP.

5. When the $I_{TPSM}$ shows positive anomalies, warm and humid water vapour from the Bay of Bengal and Arabian Sea and cold air from high latitudes merge over the central and eastern TP, coupled with lower-level convergence and upper-level divergence, which could contribute to an increase in plateau summer monsoonal precipitation. When the $I_{TPSM}$ exhibits negative anomalies, the anomalous cyclonic vapour transportation circulation to the south of the TP hinders the warm and humid air to the TP, and the anomalous anticyclonic vapour transportation circulation to the southeast of Lake Baikal weakens cold air to the south, coupled with weakened convergence and upward motion, all these lead to a decrease in plateau summer monsoonal precipitation.

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