Intraseasonal variability of Tibetan Plateau snow cover

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Abstract
Using the daily snow cover data at 24-km resolution from the Interactive Multi-sensor Snow and Ice Mapping System snow cover analysis, this study describes the variability in Tibetan Plateau (TP) snow cover (TPSC) at multiple time scales with a focus on the intraseasonal time scale (10–90 days). TPSC demonstrates variability over a wide range of temporal scales, but the annual cycle is generally dominant. Synoptic-scale variability, seasonal variability and interannual and long-term changes make small contributions to the total daily variability in TPSC. Intraseasonal variability (ISV) is dominant over most of the central and eastern TP and explains 22–40% of the total variability and leads to obvious variations in TPSC over periods shorter than a season. The ISV of TPSC is more active in the cold season than in the warm season. Specifically, the ISV over the Changtang Plateau explains approximately 50% of the total variability of snow cover in the cold season and is even more dominant than the annual cycle. Possible influences of regional atmospheric circulations on TPSC are also examined. TPSC variability is highly correlated with regional surface air temperature (SAT) and precipitation at an intraseasonal time scale. TPSC and SAT tend to have a simultaneous relationship, while anomalous precipitation leads to subsequent TPSC variations with a lag of approximately 5 days and a positive relationship. Such relationships are the result of intraseasonal variations in regional atmospheric circulation. The anomalous adiabatic heating induced by vertical ascending motion leads to tropospheric temperature variations. Furthermore, the horizontal advection of moisture and apparent moisture sink, which are induced by anomalous moisture supply and snow evaporation anomalies, respectively, lead to anomalous moisture associated with changes in the TPSC.

KEYWORDS
climatology, intraseasonal variability, multiple time scales, snow cover, Tibetan Plateau

1 INTRODUCTION
Snow cover is a crucial component in both the climate system and the cryosphere. The presence of snow cover significantly influences the ground thermal regime (Zhang, 2005). Because snow cover acts as the lower boundary condition of the atmosphere, it forces the regional and global atmosphere and can serve as an
indicator of atmospheric circulation and climate variations (Hahn and Shukla, 1976; Dickson, 1984; Barnett et al., 1989; Bamzai and Shukla, 1999; Fasullo, 2004; Wu and Kirtman, 2007; Wu et al., 2009; Allen and Zender, 2011; Henderson et al., 2018). Snow cover affects the local and global climate systems, mainly via the sensitivity of the radiation balance to the high albedo and emissivity characteristics of snow (Dery and Brown, 2007; Ghatak et al., 2014; Wu et al., 2014; Wu and Chen, 2016), the energy allocation involved in the melting of snowpack (Cohen and Rind, 1991; Kripalani et al., 2002) and snow-hydrological effects (Yasunari et al., 1991; Wu et al., 2014; Diro et al., 2018). As the source of water for streamflow and groundwater recharge, snow cover also plays an important role in hydrological cycle (Stewart et al., 2004; Rauscher et al., 2008; Koster et al., 2010; Mahanama et al., 2012; Barnhart et al., 2016).

Due to the significant impacts of snow cover, it is important to investigate the climatological characteristics and origins of snow cover variations. Generally, the snow covers of Northern Hemisphere (NH) lands at high latitude change seasonally in accordance with the solar radiation (Robinson et al., 1995). In addition to seasonal variability, snow cover is subject to interannual and decadal fluctuations (Cayan, 1996; Zhang et al., 2004; Qin et al., 2006; Brown and Robinson, 2011; Shen et al., 2011). The long-term trend of snow cover under global warming (Mote et al., 2005; Brown and Mote, 2009; McCabe and Wolock, 2010; López-Moreno et al., 2013) and its future projections (Barnett et al., 2005; Rauscher et al., 2008; Fyfe et al., 2017; Musselman et al., 2017) have also drawn many attentions. On the hemispheric scale, snow cover changes slowly with a period longer than a season. However, snow cover can vary rapidly within a season over discontinuous or sporadic permafrost zones (Wang et al., 2015; Li et al., 2018; Suriano and Leathers, 2018; Song et al., 2019; Song and Wu, 2019). Understanding the rapid variations in snow cover is important for short- and medium-range weather forecasting applications (Clark and Serreze, 2000; Li et al., 2018; Zhang et al., 2019).

The Tibetan Plateau (TP) is the highest plateau in the world and is known as the third pole. Due to its high elevation, the TP is colder and has much more snow cover than other surface regions at the same latitude. Thermal forcing by the TP influences weather and climate (Kuo and Qian, 1981; Yanai and Wu, 2006). The variability of TP snow cover (TPSC), which influences land surface thermal conditions (Chen et al., 2017) and thus influences general circulations and monsoon systems over eastern and Southern Asia, can act as an indicator of the weather and climate in and around the TP (Qian et al., 2003; Wu and Qian, 2003; Zhang et al., 2004; Zhao and Moore, 2004; Zhao et al., 2007; Lin and Wu, 2011; Wu et al., 2012; Wang et al., 2017; Lu et al., 2018). The TP is also considered the Asian water tower, and TPSC is a vital water source in Asia. Rivers including the Yangtze River, Yellow River, Yarlung Zangbo River and Mekong River have headwaters over the TP. Thus, studies on the variability of TPSC are critical for water management in downstream regions (Immerzeel et al., 2009; Zhang et al., 2012, 2013). Therefore, it is important to understand the features and mechanisms of variations in the TPSC.

Studies have aimed at quantifying the seasonal, interannual and decadal variability of TP snowpack, including the features, climate impacts and factors that control the TPSC at these time scales (e.g., Shaman and Tziperman, 2005; Pu et al., 2007; Lü et al., 2008; Ding et al., 2009; Yuan et al., 2009; Mao, 2010; Si and Ding, 2013; Zhu et al., 2015; Xiao and Duan, 2016; Basang et al., 2017; Liu et al., 2017; Wang et al., 2018a; 2019; Qiu et al., 2019). The long-term variations in TPSC under global warming have also been widely studied (Kang et al., 2010; You et al., 2011; Chen et al., 2017; Huang et al., 2017; Xu et al., 2017; Wang et al., 2018b).

Reliable information on the spatial and temporal variability of snow cover is important for climate monitoring. The TP is a challenging region for snow-related research. Due to the bitter natural geographical environment and sparse population, in situ observations of snow cover are relatively lacking. Nevertheless, a number of meteorological stations over the TP region are operated by the Chinese Meteorological Administration (CMA). The in situ observations of snow cover can be directly used to research variations in the TPSC (e.g., Lü et al., 2008; Ding et al., 2009; Mao, 2010; You et al., 2011; Xu et al., 2017) and its climate effects (e.g., Qian et al., 2003; Wu and Qian, 2003; Zhang et al., 2004; Zhao et al., 2007; Zhu et al., 2015). Since most of the CMA stations are located in inhabited valleys below 4,000 m in the Southeast TP and in situ observations are used to monitor variation in the snow cover at local scales, the representativeness of this in situ data for the TP as a whole is questionable. In the absence of a large, distributed network of meteorological stations, remote sensing becomes a necessary technique. Advances in satellite remote sensing provide invaluable information on the presence of TPSC (Yang et al., 2015; Basang et al., 2017; Orsolini et al., 2019). Remote sensing data on TPSC can be evaluated against in situ station observations (Pu et al., 2007; Basang et al., 2017; Dai et al., 2017; Hao et al., 2019). Some evaluations show that remote sensing data can reliably capture the general variability of the TPSC (Yang et al., 2015; Li et al., 2018).

From perspective of the seasonal mean, the snow-covered area over the TP is established in autumn and persists to the following spring (Pu et al., 2007), and even to summer in the western and southeastern parts at high
elevation (Wang et al., 2018a). Notably, TPSC is unique compared with snow cover over other mid-latitude regions or higher latitudes. The TPSC is distinctly shallow, patchy and frequently in short duration (Robinson et al., 1995; Qin et al., 2006). Such unique characteristics may lead to fast variations in the TPSC within a seasonal period. Li et al. (2016; 2018) and Song et al. (2019) noticed that the intraseasonal variability (ISV) of TPSC should not be ignored and is associated with atmospheric intraseasonal oscillations. Recent studies have shown that accurate snow cover initialization can improve subseasonal and seasonal forecasts/simulations (Jeong et al., 2013; Lin et al., 2016; Senan et al., 2016), implying that snow cover is a potential indicator for forecasts at shorter time scales. However, the relatively fast ISV of TPSC has received relatively less attention than the well-recognized interannual and long-term changes in TPSC. A better understanding of the variability in TPSC at multiple time scales, including intraseasonal time scales, allows for us to understand all aspects of TPSC variability. To the authors’ knowledge, few papers have quantified the ISV of TPSC. The relationship between snow cover and local atmospheric variations over the TP at an intraseasonal time scale is also not clear. This study addresses the above issues.

The objective of this research is to understand the spatial and temporal distribution of TPSC variability at the intraseasonal time scale and the direct linkage of this variability with regional atmospheric circulation. The present study aim to increase attention on and our knowledge of the ISV in TPSC. Details on the data set used in this study are described in Section 2; some climatological aspects of TPSC are presented in Section 3; TPSC variability at multiple time scales, with a focus on the ISV, is described in Section 4; the linkage between TPSC and regional atmospheric circulations is further examined to elucidate the direct factors impacting the ISV of TPSC in Section 5; a conclusion and discussion are presented in Section 6.

2 | DATA

2.1 | Snow cover data

Daily snow cover data at 24-km resolution were obtained from the Interactive Multi-sensor Snow and Ice Mapping System (IMS) snow cover analysis (Helfrich et al., 2007) provided by the National Oceanic and Atmospheric Administration (NOAA). The IMS examines satellite images and other sources of data on snow cover and generates maps of snow cover distribution. Visible and infrared spectral data from the Polar Operational Environmental Satellites (POES) and Geostationary Orbiting Environmental Satellites were primarily used to generate snow cover data. Moderate Resolution Imaging Spectrometer (MODIS) imagery was also used for this purpose. In addition, ground weather observations from many countries were used as data in the present study. Since visible and infrared data suffer from persistent cloud cover, which makes observations difficult, microwave products from Special Sensor Microwave Imager (SSMI) and Advanced Microwave Scanning Radiometer for EOS (AMSR-E) were used in the IMS product. The IMS system also included the model output from the Snow Data Assimilation System (SNODAS) and station-mapped products. The spatial resolutions of visible, infrared, microwave and SNODAS products used in the IMS System vary from 1 to 40 km. The IMS product was manually created by NOAA NESDIS (the National Environmental Satellite, Data, and Information Service) satellite-product group analysts evaluating all available satellite imagery, automated snow mapping algorithms and other ancillary data. The IMS analysts use these multiple sources with different spatial resolutions within the interactive multisensor snow mapping system and re-grid it to map snow at a 24-km spatial resolution. The analyst begins charting using the map from the previous day, then uses the satellite inputs accordingly. The IMS system allows for faster processing times to produce snow cover maps from satellite remote sensing data. The raw snow cover analysis data were binary: 1 indicated a grid point covered by snow, and 0 indicated a nonsnow-covered grid point. Chen et al. (2012) validated the IMS snow cover analysis by a comparison with ground-based measurements over the continental United States. They found that the IMS maps demonstrate a good correspondence with the ground-based measurements. Furthermore, they suggested that, when mapping snow cover, IMS analysts use the same technique and similar sources of data (e.g., satellite imagery, in situ data and automated snow remote sensing products) over the whole NH. Therefore, it is reasonable to assume that the accuracy of snow cover mapping over the mid-latitude region of Eurasia is similar to that over North America. Yang et al. (2015) and Li et al. (2018) found that the overall accuracy of IMS snow cover analysis is greater than 90% compared with station observations over the TP. Li et al. (2018) also showed that the IMS snow cover analysis can capture the general subseasonal variability of the TPSC well by validation using daily in situ data. See Li et al. (2018) for details about the validation. The snow cover data used in this study span from March 1997 to August 2017.

2.2 | TPSC index

To measure the regional variability in TPSC, a TPSC index (TPSCI) for the entire TP was defined; this index
represents the percentage of the snow-covered area over the TP. The grid points over the TP were defined as points at an altitude of greater than 3,000 m and within 26–41°N and 70–105°E. According to these criteria, there were 7,289 grid points over the TP. The TPSCI for the entire TP was calculated from the IMS snow cover analysis and was defined as follows:

$$\text{TPSCI} = \frac{1}{n} \sum_{i=1}^{n} x_i \times 100\%,$$

where $x_i$ is the IMS snow cover analysis over the TP. If one grid point is covered by snow, $x_i = 1$; otherwise, $x_i = 0$. The unit of TPSCI is %.

In view of the remarkable regional characteristics of TPSC described below, we defined five regions to examine and compare the variabilities among different subregions of the TP. Figure 2a defines the five subregions, labelled A (30°–40°N, 70°–79°E), B (33°–38°N, 80°–92°E), C (28°–33°N, 80°–92°E), D (28.5°–40°N, 92.5°–104°E, excluding the area within 28.5°–32.5°N, 92.5°–98°E) and E (28.5°–32.5°N, 92.5°–98°E). Only areas with altitudes greater than 3,000 m were included. These subregions of the TP were chosen based on different climatology and variability characteristics, as described in Sections 3.2 and 3.3. Region A mainly includes the Pamirs and Karakoram. Region B is mainly over the Changtang Plateau (Northern Plateau). Region C includes the Kailas Range and western part of the Nyainqêntanglha Mountains. Region D mainly includes the Bayan Har Mountains and Hengduan Mountains. Region E includes areas over the eastern part of the Nyainqêntanglha Mountains. Similar to the TPSCI for the entire TP, the regional TPSCI was calculated for each subregion of the TP by averaging the IMS data over each region.

### 2.3 Atmospheric reanalysis and precipitation data

Daily averaged large-scale atmospheric reanalysis data were obtained from ERA-Interim (Dee et al., 2011). Variables, including surface air temperature (SAT), surface pressure and snow evaporation, were utilized in the analysis. The vertical integral of water vapour and its flux were directly obtained from ERA-Interim. Tropospheric air temperature, specific humidity, horizontal wind and vertical velocity at 11 pressure levels (650, 600, 550, 500, 450, 400, 350, 300, 250, 225 and 200 hPa) were also utilized in the analysis. Daily averaging was performed by using data from four times per day, except for snow evaporation, which is a forecast in the reanalysis. For the variable of snow evaporation, forecast field forecasts beginning at 0000 and 1200 UTC with a step of 3, 6, 9 and 12 hr were used for daily averaging. The reanalysis horizontal resolution was 0.5° × 0.5°. Bao and Zhang (2012) evaluated ERA-Interim reanalysis products with independent sounding observations over the TP and showed that the horizontal wind, temperature and relative humidity of ERA-Interim are highly correlated with the sounding observations at different vertical levels. Precipitation data were obtained from the Climate Prediction Center (CPC) Unified Gauge-Based Analysis of Global Daily Precipitation provided by NOAA (Xie et al., 2007). The horizontal resolution of this precipitation data was 0.5° × 0.5°. The atmospheric reanalysis and precipitation data were used to analyse snow cover from March 1997 to August 2017.

To further reduce the uncertainty due to reanalysis products in this study, we also utilized another state-of-the-art reanalysis data set, JRA-55 reanalysis (Kobayashi et al., 2015), which is independent from the ERA-Interim and CPC Precipitation data. The JRA-55 has a 6-hourly temporal resolution and a horizontal resolution of 1.25° × 1.25°. Because JRA-55 reanalysis does not provide snow evaporation information, we used the sublimation (evaporation from snow) from NCEP Climate Forecast System 6-hourly products with a horizontal resolution of 0.5° × 0.5° instead.

### 3 SOME CLIMATOLOGICAL ASPECTS

The contrasts in climatology and total variability of TPSC between the cold season and warm season are discussed in this section. The results in this section form the basis for the discussion of the ISV of TPSC in the following sections.

#### 3.1 Cold season versus warm season

Figure 1 shows the monthly climatology of TPSCI, which represents the percentage of the snow-covered area over the TP. TPSC has a robust annual climatological cycle. The climatology of TPSC is very different between winter and summer. The climatology peaks in February (approximately 41%) and is lowest in August (approximately 4%). The six largest monthly climatology values are from November to April in the next year (ranging from 32 to 41%) and the six smallest monthly climatology values are from May to October (ranging from 4 to 26%). Based on this aspect of seasonal variations, we defined the cold season as from November to April in the next year and the warm season as from May to October. Climatologically, we concluded that the TP is covered by snow for 37% and
14% of the time in the cold and warm seasons, respectively.

**Figure 1** Climatological mean of the TPSCI for each month. The values represent multiple-year averages of the percentage of snow-covered areas over the TP for each month (unit: %)

**3.2 | Spatial pattern of climatology**

Figure 2a,b shows the climatological mean of TPSC for both the cold season and warm season. The spatial pattern of climatology is nonuniform and region-dependent. In view of the unique regional characteristics of the TP, we defined five TP subregions and examined and compared their climatology (see Section 2.2 for definitions). Generally, in both the cold season and warm season, the climatology of regions A and E shows a larger occurrence of TPSC than that in Regions B–D. Most grids in Regions A and E have climatological snow-cover occurrences greater than 75% in the cold season (Figure 2a). The climatological snow-cover occurrence in Regions A and E decreases in the warm season for most grids (Figure 2b) but is greater than 25% and still larger than that in Regions B–D. The climatology in Regions B–D also decreases from the cold season to the warm season. The regionally averaged climatological mean conforms to

**Figure 2** Climatological mean of the TPSC for the (a) cold season and (b) warm season. The values represent the climatological probability of snow cover in each grid (unit: %). The boxes marked by A–B in (a) show the subregions of the TP defined in this paper. (c–d) as (a–b) but for SDs
these results (Figure 3a). The regionally averaged climatological mean in Regions A and E is more than twice that in Regions B–D for both the warm and cold seasons. The climatology in all regions decreases from the cold season to the warm season. The regionally averaged climatological mean in the cold season is twice that in the warm season for all regions.

3.3 Spatial distribution of total variability

The total variability in this study is based on daily data. The distribution of total variability is shown in Figure 2c, d. Some parts of Regions A and E have standard deviations (SDs) of less than 20% during the cold season (Figure 2c) and larger SDs during the warm season (greater than 35%). This characteristic is more visually evident for the regionally averaged snow-covered percentage (Figure 3b). SDs for Regions A and E are smaller in the cold season than those in the warm season. The situation is opposite for Regions B–D: SDs for Regions B–D are larger in the cold season than those in the warm season (Figure 2c). Regional averaging also indicates this characteristic (Figure 3b).

The high climatological mean of snow cover (greater than 60%) in Regions A and E during wintertime, as revealed in Figures 2a and 3a, leads to smaller SDs in these regions than those in the other regions because A and E are covered by snow during almost all cold seasons. The moderate climatological mean of the wintertime snow cover in Regions B–D (~30%) results in relatively large SDs. However, the situation in the warm season is quite different. Regions A and E have larger SDs than those of other regions. The climatological mean in Regions A and E during the warm season is moderate (~30%, similar to that in Regions B–D during the cold season). The decreased climatological mean snow cover in Regions A and E during the warm season leads to relatively large SDs. However, the climatological mean snow cover in Regions B–D during the warm season is too small (~10%). This small climatological probability of snow coverage induces small deviations in Regions B–D during the warm season. Briefly, a high daily SD tends to occur in regions with a moderate climatological probability of snow coverage (e.g., 30%). Either a large (e.g., 60%) or a small (e.g., 10%) climatology snow coverage probability decreases the SD.

4 MULTIPLE TIME SCALE VARIABILITY

In Section 3, we found that the climatology and total variability of TPSC differ between the cold and warm seasons and show spatial heterogeneities. In this section, characteristics of the multiple time scale variability of TPSC in both the cold and warm seasons for each subregion are presented with a focus on the ISV.

4.1 Decomposition of the total variability

We decomposed the TPSCIs for each region into four bands by filtering (Figure S1–S5). The Lanczos filters (Duchon, 1979) were applied. These four bands are synoptic-scale bands (with frequencies faster than $10^{-1}$ day$^{-1}$), intraseasonal bands (with frequencies between $10^{-1}$ and $90^{-1}$ day$^{-1}$), seasonal component bands (with frequencies between $90^{-1}$ day$^{-1}$ and $0.5^{-1}$ year$^{-1}$), and annual cycle (with frequencies between $0.5^{-1}$ and $1.5^{-1}$ year$^{-1}$) and
interannual and long-term bands (with frequencies slower than $1.5^{-1}$ year$^{-1}$). We standardized the raw TPSCIs for each region. The SD of the raw TPSCIs, which represents total variability, is 1.0. The SD of each component reflects the variability in each band. The explained variance represents how much of the total variability can be explained by each component. The SD and explained variances of each component for all regions are summarized in Table 1.

For Region A, the annual cycle is the most dominant component and explains 90.0% of the total variability (Table 1 and Figure S1). All other components are much smaller than the annual cycle component. For Region B, the intraseasonal component is the most dominant component and explains 40.0% of the total variability. The intraseasonal component is even more dominant than the annual cycle, which explains 35.7% of the total variability. Apart from the intraseasonal component and annual cycle, the other components make a relatively small contribution to the total variability. For Regions C and D, the annual cycle is the most dominant component and explains 51.5% and 61.5% of the total variability, respectively. The intraseasonal components of C and D are secondary dominant components; the intraseasonal component explains 29.1% and 21.9%, respectively, of the total variability. Apart from the intraseasonal component and annual cycle, the other components make a relatively small contribution to the total variability. For Region E, the annual cycle is the most dominant component and explains 76.6% of the total variability. Other components make a relatively small contribution to the total variability.

Similarities among the variabilities of TPSC at multiple time scales were found: (a) the synoptic-scale component and seasonal component make rather small contributions to the total variability (ranging from 0.6 to 9.9%). (b) The annual cycle is dominant (ranging from 35.7 to 90.0%). (c) Interannual and long-term components are not dominant from the perspective of daily resolution (ranging from 6.7 to 12.1%). However, unique characteristics were also identified in the regions: the intraseasonal components are dominant in Regions B–D (range from 21.9 to 40.0%), especially in Region B. However, the intraseasonal component makes relatively small contributions to the total variability in Regions A and E (4.0% and 9.6%, respectively). These results demonstrate that the ISV of TPSC over Regions B–D is dominant and a non-negligible component of the total variations.

We used the snow cover over Region B in the cold season of 2014/2015 as an example to show the snow cover variability at multiple time scales (Figure 4). The red line, dashed blue line and purple line show the intraseasonal (10–90 day) and interannual and long-term components of snow cover, respectively. The unit is %

### Table 1

|                     | Region A | Region B | Region C | Region D | Region E |
|---------------------|----------|----------|----------|----------|----------|
| Synoptic-scale (<10 days) | 0.6%     | 9.9%     | 6.5%     | 5.4%     | 2.0%     |
| Intraseasonal (10–90 days) | 4.0%     | 40.0%    | 29.1%    | 21.9%    | 9.6%     |
| Seasonal (90 days–0.5 year) | 3.8%     | 14.0%    | 10.3%    | 9.2%     | 11.5%    |
| Annual cycle (0.5–1.5 year) | 90.0%    | 35.7%    | 51.5%    | 61.5%    | 76.6%    |
| Interannual and long-term (>1.5 years) | 9.9%     | 6.7%     | 10.8%    | 9.1%     | 12.1%    |

Note: See manuscript for more detail.
the TPSC is sometimes above-normal (e.g., mid-January 2015 and early-March 2015) and sometime below-normal (e.g., mid-November 2014 and mid-February 2015) from climatological background. The intraseasonal component leads to obvious variations with a period shorter than a season, making TPSC exhibit a distinct lack of persistence within one season.

4.2 Intraseasonal variability

Section 4.1 describes the decomposition signal of TPSCI and the variability of specific bands and reveals that TPSC demonstrates variability at a wide range of time scales. Figure 5a shows the SD of the ISV of TPSC for each region during the cold and warm seasons, and Figure 5b shows the percentage of total variability that can be explained by the ISV variability. In all regions, the ISV and ISV-explained variance in the cold season are larger than those in the warm season, which is not the case for total variability (Figure 3b). This suggests that the ISV of TPSC is more active in the cold season than that in the warm season. As shown in Figure 5a, the contrasts of the ISV between the cold and warm seasons are much more drastic for Regions B–D (the ISV in the cold season is approximately twice that in the warm season) than for the other regions. The ISV-explained variance in the cold season is also larger than that in the warm season for all regions (Figure 5b), suggesting that ISV contributes to a higher proportion of the total variability in the cold season than in the warm season. Specifically, ISV explains approximately 50% of the total variability of snow cover in the cold season for Regions B–D. The ISV-explained variance for Regions A and E in the cold season and for Regions B–D in the warm season are smaller but still considerable (approximately 20–30%).

5 LINKAGE WITH REGIONAL ATMOSPHERIC CIRCULATION

The results in Section 4 indicate that the ISV of TPSC is significant across Regions B–D (mainly the central and eastern TP). Large-scale snow cover variability is largely determined by atmospheric circulation (e.g., Lü et al., 2008; Mao, 2010; Li et al., 2016; Xu et al., 2017; Wang et al., 2019). To better understand the cause of the ISV of TPSC, the linkage between regional atmospheric circulation and TPSC is analysed in this section.

5.1 Lead-lag linkage with SAT and precipitation

Temperature and precipitation determine the variability of TPSC at interannual, decadal and long-term time scales (Singh et al., 2014; Huang et al., 2017; Xu et al., 2017; Wang et al., 2018b). Meteorological variables, such as air temperature and precipitation, not only will change gradually in the future but also have short-term variability at the intraseasonal time scale, which may contribute to the ISV of TPSC. Figure 6 presents the lead-lag correlation coefficients for TPSCI and regionally averaged SAT across the central and western TP (27°–40°N, 80°–105°E), indicating a close relationship between TPSC and local SAT; the relationship between TPSC and local precipitation is also distinct. The correlation coefficients show a clear negative relationship between TPSC and SAT, which tends to be a simultaneous relationship (with an
approximately 0- or 1-day lead). This relationship is slightly stronger in the cold season than that in the warm season (compare the orange line in Figure 6a to that in Figure 6b). The relationship between TPSC and precipitation is also significant but shows a lead-lag characteristic; anomalous precipitation has a positive relationship with TPSCI and leads to TPSCI variations after approximately 5 days. The observed lead-lag relationship between TPSC and precipitation occurs because of the cumulative effect. Specifically, this relationship is much stronger in the cold season than in the warm season (compare the dashed blue line in Figure 6a to that in Figure 6b), which may be explained by seasonal changes in precipitation from snowfall to rainfall. The empirical relationship suggests that SAT and precipitation are potential factors governing TPSC variation at intraseasonal time scales.

5.2 Linkage with tropospheric temperature

We have shown that the negative relationship between TPSC and SAT tends to be simultaneous (lag = 0-day; Figure 6). The air temperature signature associated with TPSC is stronger than the near-surface temperature signature associated with TPSC. Figure 7a illustrates the correlation between TPSC and tropospheric temperature at a lag of 0 days. The results show that TPSC is associated with atmospheric temperature from the middle to upper troposphere. Note that the land surface pressure over the TP is approximately 600 hPa. This means that TPSC is linked with atmospheric temperature from the land surface to 300 hPa. The correlation coefficients are significant from the near-surface to 300 hPa and are larger in the cold season than in the warm season.

To elucidate the physical processes contributing to the association of tropospheric temperature variability with TPSC, the temperature budget equation was analysed. The temperature change $\frac{\partial T}{\partial t}$ at each constant pressure level can be written as follows:

$$\frac{\partial T}{\partial t} = -\left(u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y}\right) - \omega \left(\frac{\partial T}{\partial p} - \frac{\alpha}{C_p}\right) + \frac{Q_1}{C_p}, \quad (2)$$

where $T$ is air temperature, $t$ is time, $u$ and $v$ represent the zonal and meridional wind components, $\omega$ is vertical velocity, $p$ is pressure, $C_p$ is the specific heat at constant pressure, $\alpha$ is the specific volume and $Q_1$ indicates the atmospheric apparent heat source (Yanai et al., 1973). As illustrated in Figure 7, TPSC is significantly linked with temperature from the land surface to 300 hPa. We vertically integrated the temperature budget equation (from the land surface to 300 hPa). To describe the relative magnitude of each budget term, a composite analysis was performed. Extreme anomalous TPSC days were selected according to TPSCI. Because the temperature budget...
equation diagnoses the change of temperature ($\partial T / \partial t$), we calculated the derivative of TPSCI with respect to time ($\Delta$TPSCI, that is, rate of change of TPSCI). $\Delta$TPSCI for day $n$ is the difference between TPSCI on day $n + 1$ and day $n - 1$. Then, the $\Delta$TPSCI was standardized. The days with standardized $\Delta$TPSCI values larger than 1.0 (less than −1.0) were regarded as extremely positive (negative) anomalous days. For the composite, there were 415 and 393 extremely positive and negative anomalous days, respectively. The averaged differences between the extremely positive and negative anomalous days were regarded as composites.

The composites for the temperature budget equation are shown in Figure 8. The results for the cold season and warm season are very similar. It is not surprising that the composite for $\partial T / \partial t$ is negative because of the negative relationship between TPSC and air temperature. The $\partial T / \partial t$ is derived from three terms in the right of the equation: horizontal advection of temperature $-\nabla \cdot V T$, adiabatic heating $\omega (\partial T / \partial p - \alpha / C_p)$, and diabatic heating $Q_1 / C_p$. The absolute value composite of $-\nabla \cdot V T$ is much smaller than the other terms, indicating that horizontal advection is not an important factor for the temperature variability associated with TPSC. The negative $\partial T / \partial t$ is mainly attributed to the $\omega (\partial T / \partial p - \alpha / C_p)$ term, which also shows a negative composite value. In contrast, the $Q_1 / C_p$ term makes the opposite contribution. As a result, part of the cooling effect induced by $\omega (\partial T / \partial p - \alpha / C_p)$ is offset by $Q_1 / C_p$, but this does not change the total cooling effect. The temperature budget equation shows that the adiabatic heating induced by vertical motion leads to an

\[ \frac{\partial T}{\partial t} = \nabla \cdot V T + \omega \left( \frac{\partial T}{\partial p} - \frac{\alpha}{C_p} \right) + \frac{Q_1}{C_p} \]
Overall cooling effect. The composites for the vertical velocity profile, which show that TPSC has a relationship with negative anomalous $\omega$ (anomalous ascending motion), conform to this inference (Figure 9). The maximum composites occur from 300 to 450 hPa. These results suggest that the adiabatic cooling induced by vertical ascending motion leads to the negative anomalous tropospheric temperature associated with an increase in TPSC. The vertical descending motion leads to the opposite change in the TPSC.

5.3 Linkage with tropospheric vapour

To understand the anomalous precipitation associated with the subsequent anomalous TPSC, the linkage between atmospheric vapour and TPSC was investigated. We showed that anomalous precipitation leads to TPSCI variations of approximately 10 to 0 days (peaks after 4–5 days, see Figure 6). Here, we use the same composite method as that in Section 5.2, but composites were performed for atmospheric vapour at a lead time of 5 days. The results are shown in Figure 7b. TPSC is associated with tropospheric vapour in the middle to upper troposphere from the land surface to 350 hPa. The correlation coefficients are significant from the near-surface to 350 hPa and are much larger in the cold season than in the warm season.

To reveal the processes responsible for the association between tropospheric vapour variability and TPSC, we conducted a vertically integrated moisture budget analysis over the central and eastern TP using the following moisture tendency equation:

$$\frac{\partial q}{\partial t} = -\left( u \frac{\partial q}{\partial x} + v \frac{\partial q}{\partial y} \right) - q \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \omega \frac{\partial q}{\partial p} \right) - \omega T \frac{\partial q}{\partial p} - \frac{Q_2}{L}, \quad (3)$$

where $q$ is the specific humidity, $t$ is the time, $u$ and $v$ represent the horizontal wind components, $\omega$ is vertical velocity, $p$ is pressure and $Q_2$ is the atmospheric apparent moisture sink (Yanai et al., 1973) that includes net latent heating and vertical convergence of eddy moisture transport and $L$ is the latent heat of condensation. Vertical integration of the moisture budget equation (from the land surface to 300 hPa) was performed. Composite analysis was performed as described in Section 5.2.

The composites for the moisture budget equation are shown in Figure 10. The composite for $\partial q/\partial t$ is positive, corresponding to the positive relationship between TPSC and precipitation. The $\partial q/\partial t$ is derived from the four terms in right of the equation: horizontal advection of moisture $-(u \frac{\partial q}{\partial x} + v \frac{\partial q}{\partial y})$, divergence of moisture $-q(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \omega \frac{\partial q}{\partial p})$, vertical advection $-\omega T \frac{\partial q}{\partial p}$ and the atmospheric apparent moisture sink $-\frac{Q_2}{L}$. Both the composites of the horizontal advection term and the atmospheric apparent moisture sink term are positive anomalies that are attributed to the anomalous $\partial q/\partial t$.

**Figure 9** Composites of the 10–90 day vertical velocity (unit: $10^{-2}$ Pa s$^{-1}$) for TPSCI. The x-axis indicates vertical velocity. The y-axis indicates pressure levels (unit: hPa). Blue lines and dashed red lines indicate values for the cold and warm seasons, respectively. All the composites are significant at the 99% confidence level.

**Figure 10** Same as Figure 8, but for the moisture budget (unit: $10^{-4}$ kg kg$^{-1}$ day$^{-1}$) at a lead of 5 days. The terms from left to right are specific humidity tendency, horizontal moisture advection, moisture divergence, vertical moisture advection and latent heating.
The divergence term is relatively small. The results for the cold season and warm season are similar, except for the vertical advection term. The vertical advection term has a relatively large negative value in the cold season but is negligible in the warm season. This may be because of the different climatological backgrounds of the cold and warm seasons. Overall, the moisture budget equation shows that the horizontal advection of moisture and the apparent moisture sink lead to anomalous moisture.

The anomalous horizontal moisture advection is caused by moisture anomalies upstream of the TP (vectors in Figure 11). Climatological westerly winds occur over the area upstream of the TP and bring water vapour from the west of the TP to the central and eastern TP, thereby forming a moisture channel (Li et al., 2016). The anomalous horizontal moisture advection then influences the regional moisture budget. The atmospheric apparent moisture sink (Yanai et al., 1973) includes net latent heating and the vertical convergence of eddy moisture transport. Here, we found that the apparent moisture sink is related to evaporation processes. The latent heating is caused by snow evaporation over the TP (shading in Figure 11). The snow evaporation anomalies related to changes in the TPSC mainly occur over the eastern TP and contribute to increased moisture.

This section describes analysis of the direct linkage of this variability with regional atmospheric circulation using ERA-interim reanalysis and CPC precipitation data. To verify the robustness of our results regardless of the reanalysis data sets, we repeated all analyses using another state-of-the-art reanalysis data set (Figure S5–S11). The results based on the JRA-55 data set are nearly identical to those of ERA-Interim, with some trivial differences. The differences raising from the uncertainties of atmospheric reanalysis data are much smaller than the variations of atmosphere related to TPSC and do not change the conclusions at all.

6 | CONCLUSION AND DISCUSSION

TPSC plays an important role in climate and hydrological systems. The present study described the variability in TPSC at multiple time scales with a focus on the intraseasonal time scale (10–90 days). Possible influences of regional atmospheric circulations on TPSC were also examined. The final conclusions are as follows:

1. Large spatial heterogeneities in the climatology and total variability of the TPSC were observed. High total variability tended to occur in regions with a moderate climatological probability of snow coverage (approximately 30%), including Regions B–D (mainly central and eastern TP) during the cold season (November–April) and Regions A (mainly the western TP) and E (eastern part of the Nyainqêntanglha Mountains) during the warm season (May–October).

2. TPSC demonstrates variability at a wide range of temporal scales. Synoptic-scale components, seasonal components and interannual and long-term components of TPSC show rather small contributions to the total variability in snow cover (less than 14.0%). The annual cycle is dominant (ranging from 35.7 to 90.0%). Intraseasonal components are dominant for Regions B–D (range from 21.9 to 40.0%). These

![Figure 11](image-url)

**Figure 11** Composites of the 10–90 day vertical integral of moisture flux (vectors; unit: kg m⁻¹ s⁻¹) and snow evaporation (shading; m of water) for TPSCI at a lead of 5 days. These composites were calculated from moisture and evaporation data leading the TPSCI tendency by 5 days. (a) and (b) represent the cold and warm seasons, respectively. Only vectors and shading with composites that are statistically significant at the 99%
intraseasonal components lead to obvious variations with a period shorter than one season, making TPSC exhibit a distinct lack of persistence within individual seasons.

3. The ISV of TPSC is more active in the cold season than in the warm season. Specifically, the ISV over Region B explained approximately 50% of the total variability in snow cover in the cold season (mainly the Changtang Plateau) and was even more dominant than the annual cycle. The ISV-explained variances for Regions A and E in the cold season and for Regions B–D in the warm season are smaller but still considerable (approximately 20–30%).

4. The ISV of TPSC is highly correlated with regional SAT and precipitation, suggesting that SAT and precipitation are potential factors governing TPSC variations at intraseasonal time scales. TPSC and SAT tend to have a simultaneous relationship, while anomalous precipitation leads to subsequent TPSC variations after approximately 5 days, with a positive relationship.

5. The direct relationships of TPSC with SAT and precipitation are the result of intraseasonal variations in regional atmospheric circulation. The adiabatic cooling induced by vertical ascending motion leads to negative anomalous tropospheric temperatures associated with an increase in TPSC, and vertical descending motion leads to the opposite change in TPSC. Moreover, the horizontal advection of moisture and apparent moisture sink, which are induced by anomalous moisture supply and snow evaporation anomalies, respectively, lead to anomalous moisture.

This article describes a general analysis of the ISV of TPSC. More detailed work is necessary in the future. Based on this study, investigations of the following subjects will be valuable in the near future. (a) Only the regional atmospheric circulation related to TPSC was discussed. TPSC may be related to atmospheric circulation outside the TP. More work is necessary to investigate the effects of changes in atmospheric circulation on TPSC. (b) We investigated the overall ISV of TPSC. More detailed studies on each subregion, the ISV component (quasi-biweekly oscillation or 30–60 day), and related atmospheric circulation variability are needed. (c) A detailed analysis of the characteristics of snow-cover melting and snow-cover building is necessary to understand the changes in regional evaporation. The key mechanisms causing the regional differences in TPSC variation also need to be elucidated; further studies of different sub-regions are required. (d) Snow-atmosphere coupling should be noted. The relationship between SAT and TPSC is simultaneous. SAT influences TPSC and may also be affected by TPSC due to snow-cover feedback (e.g., the albedo effect). Numerical experiments might isolate the impact of SAT and the feedback effect and be valuable for further investigations.

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CONFLICT OF INTEREST
The authors declare no potential conflict of interest.

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REFERENCES
Allen, R.J. and Zender, C.S. (2011) Forcing of the Arctic oscillation by Eurasian snow cover. Journal of Climate, 24, 6528–6539.
Bamzai, A.S. and Shukla, J. (1999) Relation between Eurasian snow cover, snow depth, and the Indian summer monsoon: an observational study. Journal of Climate, 12, 3117–3132.
Bao, X. and Zhang, F. (2012) Evaluation of NCEP–CFSR, NCEP–NCAR, ERA-Interim, and ERA-40 reanalysis datasets against independent sounding observations over the Tibetan Plateau. Journal of Climate, 26, 206–214.
Barnett, T.P., Adam, J.C. and Lettenmaier, D.P. (2005) Potential impacts of a warming climate on water availability in snow-dominated regions. Nature, 438, 303–309.
Barnett, T.P., Dümenil, L., Schießl, U., Roeckner, E. and Latif, M. (1989) The effect of Eurasian snow cover on regional and global climate variations. Journal of the Atmospheric Sciences, 46, 661–686.
Barnhart, T.B., Molotch, N.P., Livneh, B., Harpold, A.A., Knowles, J.F. and Schneider, D. (2016) Snowmelt rate dictates streamflow. Geophysical Research Letters, 43, 8006–8016.
Basang, D., Barthel, K. and Olseth, A.J. (2017) Satellite and ground observations of snow cover in Tibet during 2001–2015. Remote Sensing, 9, 1201.
Brown, R.D. and Mote, P.W. (2009) The response of Northern Hemisphere snow cover to a changing climate. *Journal of Climate*, 22, 2124–2145.

Brown, R.D. and Robinson, D.A. (2011) Northern Hemisphere spring snow cover variability and change over 1922-2010 including an assessment of uncertainty. *The Cryosphere*, 5, 219–229.

Cayan, D.R. (1996) Interannual climate variability and snowpack in the western United States. *Journal of Climate*, 9, 928–948.

Chen, C., Lakhankar, T., Romanov, P., Helfrich, S., Powell, A. and Khanbilvardi, R. (2012) Validation of NOAA-Interactive Multisensor Snow and Ice Mapping System (IMS) by comparison with ground-based measurements over continental United States. *Remote Sensing*, 4, 1134–1145.

Chen, X.N., Long, D., Hong, Y., Liang, S.L. and Hou, A.Z. (2017) Observed radiative cooling over the Tibetan Plateau for the past three decades driven by snow cover-induced surface albedo anomaly. *Journal of Geophysical Research-Atmospheres*, 122, 6170–6185.

Clark, M.P. and Serreze, M.C. (2000) Effects of variations in east Asian snow cover on modulating atmospheric circulation over the North Pacific Ocean. *Journal of Climate*, 13, 3700–3710.

Cohen, J. and Rind, D. (1991) The effect of snow cover on the climate. *Journal of Climate*, 4, 689–706.

Dai, L., Che, T., Ding, Y. and Hao, X. (2017) Evaluation of snow cover and snow depth on the Qinghai-Tibetan Plateau derived from passive microwave remote sensing. *The Cryosphere*, 11, 1933–1948.

Dee, D.P., Uppala, S.M., Simmons, A.J., Berrisford, P., Poli, P., Kobayashi, S., Andrae, U., Balmaseda, M.A., Balsamo, G., Bauer, P., Bechtold, P., Beljaars, A.C.M., van de Berg, L., Bidlot, J., Bormann, N., Delsol, C., Dragani, R., Fuentes, M., Geer, A.J., Haimberger, L., Healy, S.B., Hersbach, H., Holm, E. V., Isaksen, L., Kallberg, P., Koehler, M., Matricardi, M., McNally, A.P., Monge-Sanz, B.M., Morcrette, J.J., Park, B.K., Peubey, C., de Rosnay, P., Tavolato, C., Thepaut, J.N. and Vitart, F. (2011) The ERA-Interim reanalysis: configuration and performance of the data assimilation system. *Quarterly Journal of the Royal Meteorological Society*, 137, 553–597.

Dery, S.J. and Brown, R.D. (2007) Recent Northern Hemisphere snow cover extent trends and implications for the snow-albedo feedback. *Geophysical Research Letters*, 34, L22504.

Dickson, R.R. (1984) Eurasian snow cover versus Indian monsoon rainfall: an extension of the Hahn-Shukla results. *Journal of Climate and Applied Meteorology*, 23, 171–173.

Ding, Y., Sun, Y., Wang, Z., Zhu, Y. and Song, Y. (2009) Inter-decadal variation of the summer precipitation in China and its association with decreasing Asian summer monsoon part II: possible causes. *International Journal of Climatology*, 29, 1926–1944.

Diro, G.T., Sushima, L. and Huzyi, O. (2018) Snow-atmosphere coupling and its impact on temperature variability and extremes over North America. *Climate Dynamics*, 50, 2993–3007.

Duchon, C.E. (1979) Lanczos filtering in one and two dimensions. *Journal of Applied Meteorology*, 18, 1016–1022.

Fasullo, J. (2004) A stratified diagnosis of the Indian monsoon-Eurasian snow cover relationship. *Journal of Climate*, 17, 1110–1122.

Fye, J.C., Derksen, C., Mudryk, L., Flato, G.M., Santer, B.D., Swart, N.C., Molotch, N.P., Zhang, X., Wan, H., Arora, V.K., Scinocca, J. and Jiao, Y. (2017) Large near-term projected snowpack loss over the western United States. *Nature Communications*, 8, 14966.

Ghatak, D., Sinsky, E. and Miller, J. (2014) Role of snow-albedo feedback in higher elevation warming over the Himalayas, Tibetan Plateau and Central Asia. *Environmental Research Letters*, 9, 114008.

Hahn, D.G. and Shukla, J. (1976) An apparent relationship between Eurasian snow cover and Indian monsoon rainfall. *Journal of the Atmospheric Sciences*, 33, 2461–2462.

Hao, X., Luo, S., Che, T., Wang, J., Li, H., Dai, L., Huang, X. and Feng, Q. (2019) Accuracy assessment of four cloud-free snow cover products over the Qinghai-Tibetan Plateau. *International Journal of Digital Earth*, 12, 375–393.

Helfrich, S.R., McNamara, D., Ramsay, B.H., Baldwin, T. and Kasheta, T. (2007) Enhancements to, and forthcoming developments in the Interactive Multisensor Snow and Ice Mapping System (IMS). *Hydrological Processes*, 21, 1576–1586.

Henderson, G.R., Peings, Y., Furtado, J.C. and Kushner, P.J. (2018) Snow–atmosphere coupling in the Northern Hemisphere. *Nature Climate Change*, 8, 954–963.

Huang, X.D., Deng, J., Wang, W., Feng, Q.S. and Liang, T.G. (2017) Impact of climate and elevation on snow cover using integrated remote sensing snow products in Tibetan Plateau. *Remote Sensing of Environment*, 190, 274–288.

Immerzeel, W.W., Droogers, P., de Jong, S.M. and Bierkens, M.F.P. (2009) Large-scale monitoring of snow cover and runoff simulation in Himalayan river basins using remote sensing. *Remote Sensing of Environment*, 113, 40–49.

Jeong, J.H., Linderholm, H.W., Woo, S.H., Folland, C., Kim, B.M., Kim, S.J. and Chen, D.L. (2013) Impacts of snow initialization on seasonal forecasts of surface air temperature for the cold season. *Journal of Climate*, 26, 1956–1972.

Kang, S.C., Xu, Y.W., You, Q.L., Flugel, W.A., Pepin, N. and Yao, T.D. (2010) Review of climate and cryospheric change in the Tibetan Plateau. *Environmental Research Letters*, 5, 015011.

Kobayashi, S., Ota, Y., Harada, Y., Ebitu, A., Moriya, M., Onoda, H., Onogi, K., Kamahori, H., Kobayashi, C., Endo, H., Miyaoaka, K. and Takahashi, K. (2015) The JRA-55 reanalysis: general specifications and basic characteristics. *Journal of the Meteorological Society of Japan. Ser. II*, 93, 5–48.

Koster, R.D., Mahanama, S.P.P., Livneh, B., Lettenmaier, D.P. and Reichele, R.H. (2010) Skill in streamflow forecasts derived from large-scale estimates of soil moisture and snow. *Nature Geoscience*, 3, 613–616.

Kripalani, R.H., Kim, B.-J., Oh, J.-H. and Moon, S.-E. (2002) Relationship between Soviet snow and Korean rainfall. *International Journal of Climatology*, 22, 1313–1325.

Kuo, H.L. and Qian, Y.F. (1981) Influence of the Tibetan Plateau on cumulative and diurnal changes of weather and climate in summer. *Monthly Weather Review*, 109, 2337–2356.

Li, W., Guo, W., Hsu, P.-C. and Xue, Y. (2016) Influence of the Madden–Julian oscillation on Tibetan Plateau snow cover at the intraseasonal time-scale. *Scientific Reports*, 6, 30456.

Li, W., Guo, W., Qiu, B., Xue, Y., Hsu, P.-C. and Wei, J. (2018) Influence of Tibetan Plateau snow cover on east Asian
atmospheric circulation at medium-range time scales. Nature Communications, 9, 4243.

Lin, H. and Wu, Z. (2011) Contribution of the autumn Tibetan Plateau snow cover to seasonal prediction of North American winter temperature. Journal of Climate, 24, 2801–2813.

Lin, P., Wei, J., Yang, Z.L., Zhang, Y. and Zhang, K. (2016) Snow data assimilation-constrained land initialization improves seasonal temperature prediction. Geophysical Research Letters, 43, 11423–11432.

Liu, S.Z., Wu, Q.G., Ren, X.J., Yao, Y.H., Schroeder, S.R. and Hu, H. (2017) Modeled Northern Hemisphere autumn and winter climate responses to realistic Tibetan Plateau and Mongolia snow anomalies. Journal of Climate, 30, 9435–9454.

López-Moreno, J.I., Pomeroy, J.W., Revuelto, J. and Vicente-Serrano, S.M. (2013) Response of snow processes to climate change: spatial variability in a small basin in the Spanish Pyrenees. Hydrological Processes, 27, 2637–2650.

Lü, J.M., Ju, J.H., Kim, S.J., Ren, J.Z. and Zhu, Y.X. (2008) Arctic Oscillation and the autumn/winter snow depth over the Tibetan Plateau. Journal of Geophysical Research-Atmospheres, 113, D14117.

Lu, M.M., Yang, S., Li, Z.N., He, B., He, S. and Wang, Z.Q. (2018) Possible effect of the Tibetan Plateau on the “upstream” climate over West Asia, North Africa, South Europe and the North Atlantic. Climate Dynamics, 51, 1485–1498.

Mahanama, S., Livneh, B., Koster, R., Lettenmaier, D. and Reichle, R. (2012) Soil moisture, snow, and seasonal streamflow forecasts in the United States. Journal of Hydrometeorology, 13, 189–203.

Mao, J. (2010) Interannual variability of snow depth over the Tibetan Plateau and its associated atmospheric circulation anomalies. Atmospheric and Oceanic Science Letters, 3, 213–218.

McCabe, G.J. and Wolock, D.M. (2010) Long-term variability in Northern Hemisphere snow cover and associations with warmer winters. Climatic Change, 99, 141–153.

Mote, P.W., Hamlet, A.F., Clark, M.P. and Lettenmaier, D.P. (2005) Declining mountain snowpack in western North America. Bulletin of the American Meteorological Society, 86, 39–50.

Musselman, K.N., Clark, M.P., Liu, C., Ikeda, K. and Rasmussen, R. (2017) Slower snowmelt in a warmer world. Nature Climate Change, 7, 214–219.

Orsolini, Y., Wegmann, M., Dutra, E., Liu, B., Balsamo, G., Yang, K., de Rosnay, P., Zhu, C., Wang, W., Senan, R. and Arduini, G. (2019) Evaluation of snow depth and snow cover over the Tibetan Plateau in global reanalyses using in situ and satellite remote sensing observations. The Cryosphere, 13, 2221–2239.

Pu, Z., Xu, L. and Salomonson, V.V. (2007) MODIS/Terra observed seasonal variations of snow cover over the Tibetan Plateau. Geophysical Research Letters, 34, L06706.

Qian, Y.F., Zheng, Y.Q., Zhang, Y. and Miao, M.Q. (2003) Responses of China’s summer monsoon climate to snow anomaly over the Tibetan Plateau. International Journal of Climatology, 23, 593–613.

Qin, D.H., Liu, S.Y. and Li, P.J. (2006) Snow cover distribution, variability, and response to climate change in western China. Journal of Climate, 19, 1820–1833.

Qiu, B., Li, W., Wang, X., Shang, L., Song, C., Guo, W. and Zhang, Y. (2019) Satellite-observed solar-induced chlorophyll fluorescence reveals higher sensitivity of alpine ecosystems to snow cover on the Tibetan Plateau. Agricultural and Forest Meteorology, 271, 126–134.

Rauscher, S.A., Pal, J.S., Diffenbaugh, N.S. and Benedetti, M.M. (2008) Future changes in snowmelt-driven runoff timing over the western US. Geophysical Research Letters, 35, L16703.

Robinson, D.A., Frei, A. and Serreze, M.C. (1995) Recent variations and regional relationships in Northern Hemisphere snow cover. Annals of Glaciology, 21, 71–76.

Senan, R., Orsolini, Y.J., Weisheimer, A., Vitart, F., Balsamo, G., Stockdale, T.N., Dutra, E., Doblas-Reyes, F.J. and Basang, D. (2016) Impact of springtime Himalayan-Tibetan Plateau snowpack on the onset of the Indian summer monsoon in coupled seasonal forecasts. Climate Dynamics, 47, 2709–2725.

Shaman, J. and Tziperman, E. (2005) The effect of ENSO on Tibetan plateau snow depth: a stationary wave teleconnection mechanism and implications for the south Asian monsoons. Journal of Climate, 18, 2067–2079.

Shen, C.M., Wang, W.C. and Zeng, G. (2011) Decadal variability in snow cover over the Tibetan Plateau during the last two centuries. Geophysical Research Letters, 38, L10703.

Si, D. and Ding, Y.H. (2013) Decadal change in the correlation pattern between the Tibetan Plateau winter snow and the East Asian summer precipitation during 1979–2011. Journal of Climate, 26, 7622–7634.

Singh, S.K., Rathore, B.P., Bahuguna, I.M. and Ajaï. (2014) Snow cover variability in the Himalayan-Tibetan region. International Journal of Climatology, 34, 446–452.

Song, L. and Wu, R. (2019) Intrasessional snow cover variations over western Siberia and associated atmospheric processes. Journal of Geophysical Research-Atmospheres, 124, 8994–9010.

Song, L., Wu, R. and An, L. (2019) Different sources of 10–30 day intraseasonal variations of autumn snow over western and eastern Tibetan Plateau. Geophysical Research Letters, 46, 9118–9125.

Stewart, I.T., Cayan, D.R. and Dettinger, M.D. (2004) Changes in snowmelt runoff timing in western North America under a “business as usual” climate change scenario. Climatic Change, 62, 217–232.

Suriano, Z.J. and Leathers, D.J. (2018) Great lakes basin snow-cover ablation and synoptic-scale circulation. Journal of Applied Meteorology and Climatology, 57, 1497–1510.

Wang, C.H., Yang, K., Li, Y.L., Wu, D. and Bo, Y. (2017) Impacts of snow depth and timing on streamflow and runoff in a midlatitude region. Journal of Hydrometeorology, 18, 452–469.

Wang, Z., Wu, R. and Huang, G. (2018b) Low-frequency snow cover variability of the Tibetan Plateau. Climatic Change, 145, 1287–1301.

Wang, Z., Wu, R., Zhao, P., Yao, S.-L. and Jia, X. (2019) Formation of snow cover anomalies over the Tibetan Plateau in cold seasons. Journal of Geophysical Research-Atmospheres, 124, 4873–4890.
Wu, B., Yang, K. and Zhang, R. (2009) Eurasian snow cover variability and its association with summer rainfall in China. *Advances in Atmospheric Sciences*, 26, 31–44.

Wu, R. and Chen, S. (2016) Regional change in snow water equivalent–surface air temperature relationship over Eurasia during boreal spring. *Climate Dynamics*, 47, 2425–2442.

Wu, R. and Kirtman, B.P. (2007) Observed relationship of spring and summer East Asian rainfall with winter and spring Eurasian snow. *Journal of Climate*, 20, 1285–1304.

Wu, R.G., Liu, G. and Ping, Z. (2014) Contrasting Eurasian spring and summer climate anomalies associated with western and eastern Eurasian spring snow cover changes. *Journal of Geophysical Research-Atmospheres*, 119, 7410–7424.

Wu, T.W. and Qian, Z.A. (2003) The relation between the Tibetan winter snow and the Asian summer monsoon and rainfall: an observational investigation. *Journal of Climate*, 16, 2038–2051.

Wu, Z., Li, J., Jiang, Z. and Ma, T. (2012) Modulation of the Tibetan Plateau snow cover on the ENSO teleconnections: from the East Asian summer monsoon perspective. *Journal of Climate*, 25, 2481–2489.

Xiao, Z.X. and Duan, A.M. (2016) Impacts of Tibetan Plateau snow cover on the interannual variability of the East Asian summer monsoon. *Journal of Climate*, 29, 8495–8514.

Xie, P., Yatagai, A., Chen, M., Hayasaka, T., Fukushima, Y., Liu, C. and Yang, S. (2007) A gauge-based analysis of daily precipitation over East Asia. *Journal of Hydrometeorology*, 8, 607–626.

Xu, W.F., Ma, L.J., Ma, M.N., Zhang, H.C. and Yuan, W.P. (2017) Spatial-temporal variability of snow cover and depth in the Qinghai-Tibetan Plateau. *Journal of Climate*, 30, 1521–1533.

Yanai, M., Esbensen, S. and Chu, J.-H. (1973) Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets. *Journal of the Atmospheric Sciences*, 30, 611–627.

Yanai, M. and Wu, G.-X. (2006) Effects of the Tibetan Plateau. In: Wang, B. (Ed.) *The Asian Monsoon*. Berlin: Springer, pp. 513–549.

Yang, J., Jiang, L., Ménard, C.B., Luojus, K., Lemmetyinen, J. and Pulliainen, J. (2015) Evaluation of snow products over the Tibetan Plateau. *Hydrological Processes*, 29, 3247–3260.

Yasunari, T., Kitoh, A. and Tokioka, T. (1991) Local and remote responses to excessive snow mass over Eurasia appearing in the northern spring and summer climate: a study with the MRI-GCM. *Journal of the Meteorological Society of Japan. Ser. II*, 69, 473–487.

You, Q., Kang, S., Ren, G., Fraedrich, K., Pepin, N., Yan, Y. and Ma, L. (2011) Observed changes in snow depth and number of snow days in the eastern and central Tibetan Plateau. *Climate Research*, 46, 171–183.

Yuan, C., Tozuka, T., Miyasaka, T. and Yamagata, T. (2009) Respective influences of IOD and ENSO on the Tibetan snow cover in early winter. *Climate Dynamics*, 33, 509–520.

Zhang, G., Xie, H., Yao, T., Liang, T. and Kang, S. (2012) Snow cover dynamics of four lake basins over Tibetan Plateau using time series MODIS data (2001–2010). *Water Resources Research*, 48, W10529.

Zhang, L.L., Su, F.G., Yang, D.Q., Hao, Z.C. and Tong, K. (2013) Discharge regime and simulation for the upstream of major rivers over Tibetan Plateau. *Journal of Geophysical Research-Atmospheres*, 118, 8500–8518.

Zhang, T.J. (2005) Influence of the seasonal snow cover on the ground thermal regime: an overview. *Reviews of Geophysics*, 43, RG4002.

Zhang, Y., Zou, T. and Xue, Y. (2019) An Arctic-Tibetan connection on subseasonal to seasonal time scale. *Geophysical Research Letters*, 46, 2790–2799.

Zhang, Y.S., Li, T. and Wang, B. (2004) Decadal change of the spring snow depth over the Tibetan Plateau: the associated circulation and influence on the East Asian summer monsoon. *Journal of Climate*, 17, 2780–2793.

Zhao, H.X. and Moore, G.W.K. (2004) On the relationship between Tibetan snow cover, the Tibetan plateau monsoon and the Indian summer monsoon. *Geophysical Research Letters*, 31, L14204.

Zhao, P., Zhou, Z.J. and Liu, J.P. (2007) Variability of Tibetan spring snow and its associations with the hemispheric extratropical circulation and east Asian summer monsoon rainfall: an observational investigation. *Journal of Climate*, 20, 3942–3955.

Zhu, Y., Liu, H., Ding, Y., Zhang, F. and Li, W. (2015) Interdecadal variation of spring snow depth over the Tibetan Plateau and its influence on summer rainfall over East China in the recent 30 years. *International Journal of Climatology*, 35, 3654–3660.

**SUPPORTING INFORMATION**

Additional supporting information may be found online in the Supporting Information section at the end of this article.

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