Observed Large-Scale Structures and Diabatic Heating Profiles of Precipitation Over the Tibetan Plateau and South China

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Abstract This study investigates and compares large-scale dynamic and thermodynamic structures of precipitation over the Tibetan Plateau (TP) and South China (SC). Vertical velocity, temperature and moisture advection, apparent heat source $Q_1$, and apparent moisture sink $Q_2$ are analyzed for strong, moderate, and light/no-rain based on the measurements. The principal findings are as follows. (1) Both regions showed the strongest upward motion at around 400 hPa for strong and moderate rainfall, but the intensity of upward motion in SC was twice that of the TP. (2) Vertical temperature advection over the TP showed advective warming in the lower troposphere and advective cooling in the upper troposphere, but this advection over SC was weaker and had a different vertical structure. Moreover, the strong-rain horizontal temperature advection was the strongest over the TP but the weakest over SC. (3) The moisture advection over the TP reached its maximum at around 400 hPa, showing a one-peak structure; while that over SC was stronger in the mid-to-low troposphere, showing a multi-peak structure. (4) The $Q_1$ profiles of the three rain categories in the TP were top-heavy, with two heating peaks in the mid-to-upper troposphere. In contrast, the $Q_1$ profiles in SC only had a heating peak in the middle troposphere. Correlations between the vertically integrated $<Q_1>$ and $<Q_2>$ were significant in SC but weakened in the TP. (5) Diurnal cycles of large-scale structures in the TP were strong from the late afternoon to early morning, while those in SC were strong in the afternoon.

1. Introduction

The Tibetan Plateau (TP) and South China (SC) are both affected by the Asian summer monsoon, which provides sufficient energy and moisture for convection and precipitation (Chang, 2004; He et al., 2007; Johnson & Houze, 1987). However, due to the distinct geographic location and topography – the TP is situated in the west high-altitude interior while SC is located in the low-altitude southeast coast – the effect of the summer monsoon on these two regions is different. The summer monsoon usually occurs from April to September in Asia, but the beginning of the monsoon-influenced rainy season in SC is earlier than that in the TP. The summer monsoon usually arrives in SC in early-to-mid May (Chen, 1983; Ding, 1992; Ding & Chan, 2005), and then triggers strong convection and precipitation from late May to June (Ding, 1992; Luo et al., 2013; Xu et al., 2009). As for the TP, the summer monsoon prevails from June to September, but the strongest convection and precipitation occur in July and August (Fujinami & Yasunari, 2001; Qie et al., 2014; Wu et al., 2013). Most of the precipitation in the TP and SC during the active rainy season is convective precipitation (Ding, 1992; Flohn, 1968; Luo et al., 2013; Uyeda et al., 2001). However, because of the large high terrain and thin air, the precipitation in the TP is weaker than that in SC. The convective systems in SC are strong, deep, and widely distributed, whereas those in the TP are weaker, shallow, and narrowly distributed (Luo et al., 2011; Wu et al., 2013; Xu, 2013). Therefore, there is value in investigating precipitation characteristics regarding the regional variations (i.e., mountainous region vs. plain, subtropical region vs. midlatitude region, and highly monsoon-influenced region vs. less monsoon-influenced region).

Convection and precipitation processes over the TP and SC are important components of the Asian climate, which can affect large-scale circulation and thermodynamics through the release of latent heat and the redistribution of heat, moisture, and momentum (Arakawa & Schubert, 1974; Park et al., 2007; Randall et al., 1989; Sui & Yanai, 1986; Yanai & Tomita, 1998). For the TP, the large terrain effect also makes a big
difference in the transportation of energy and moisture, further affecting the circulation and precipitation systems over East Asia (Duan & Wu, 2005; Fu et al., 2006; Tao & Ding, 1981; Wu et al., 2012; Yang et al., 2014). For SC, it is a downstream region of the TP and lies on the north of the South China Sea and the west of the Pacific Ocean. Hence, the energy, water vapor, and cloud systems coming from the oceans and the plateau can move across SC and change through the convective activities (Chen et al., 2015; Ding & Chan, 2005; Xu et al., 2009). Due to these, the TP and SC regions are often considered as two important subsystems of Asian or even global climate (Ding et al., 2008; Fu et al., 2006; Lau, 1992; Wang & Linho, 2002). Studying and comparing the characteristics of precipitation between the TP and SC is helpful to understand the changes in convective systems, atmospheric environments, and weather in Asia.

There have been recent studies for convection and precipitation over these two regions based on various measurements and data sources. For example, Luo et al. (2011) used CloudSat and Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) data to analyze the frequency, vertical structure, system size, and local environment of deep convection in the monsoon regions including the TP and SC. Xu (2013) analyzed Tropical Rainfall Measuring Mission (TRMM)-based precipitation data to show that rainfall over the TP and SC significantly depended on mixed-phase microphysics processes, however, large deep convective systems contributed a much higher fraction of the total rainfall over SC than over the TP. Luo et al. (2013) found enhanced rain accumulation and convection intensity in SC during the active monsoon period using TRMM and CloudSat/CALIPSO satellite products and surface observations. Chen et al. (2015) investigated the evolutions of summer clouds and precipitation over the TP and SC using the International Satellite Cloud Climatology Project products, as well as examined the interannual variations of the diabatic heating and drying using the NCEP-NCAR reanalysis. However, most of these studies mainly concentrated on the properties of precipitation and convection (such as spatial and temporal evolutions, frequency, and intensity), the large-scale vertical dynamic and thermodynamic structures corresponding to precipitation in these two regions are less discussed.

This study is motivated by the question: how many similarities or differences in the large-scale structures and budgets during rainy periods between these two typical regions in East Asia? As mentioned above, precipitation can interact with large-scale circulation and environments through the redistribution of energy and moisture. Therefore, investigation of the association between precipitation and its large-scale dynamic structures and heat and moisture responses is conducive to help further understand the meteorological differences between the TP and SC.

The aim of this study is to compare the vertical motion, temperature advection, moisture advection, and diabatic heating and drying structures during strong, moderate, and light/no rainfall between the TP and SC in the active rainy season. The diagnostic quantities are calculated from multi-source measurements and the ERA5 reanalysis data from the European Center for Medium-Range Weather Forecasts (ECMWF). A one-dimensional constrained variational analysis (CVA) method proposed by Zhang and Lin (1997) is used in this study to process these data. The CVA method can deal with different measurements over a small region, thereby obtaining grid-scale vertical velocity and advective tendencies. Given the inevitable uncertainties in the raw measurements, this method adjusts the observed state variables by the minimum possible amount to conserve column-integrated mass, moisture, static energy, and momentum (Zhang & Lin, 1997). In this process, domain-averaged surface and top-of-the-atmosphere (TOA) observations including the surface and TOA radiative fluxes, surface latent and sensible heat fluxes, and surface precipitation are used as strong constraints. Diabatic heating and drying defined as apparent heat source $Q_1$ and apparent moisture sink $Q_2$ (Yanai et al., 1973) can be estimated as the residuals of heat and moisture budgets of large-scale motion.

Since the analysis data derived by CVA are mainly based on the measurements, they can avoid the deficiencies in physical parameterizations of simulation. Uncertainties of the analysis data are mainly from instruments and measurements, errors associated with adjustments to the state variables, and errors of the calculation in the CVA procedure. Thus, the quality of these analysis data partially relies on the accuracy of observations, especially surface precipitation (Schumacher et al., 2007; Zhang, Lin, et al., 2001). Zhang, Lin, et al. (2001) showed that the CVA method could reduce the sensitivity of the analysis data to uncertainties in the input data. Waliser et al. (2002) found that the CVA values (with mass, heat, moisture, and momentum constrained) typically exhibited smaller errors than those from the conventional objective analysis.
(only with mass constrained). Xie et al. (2003) compared the large-scale forcing data derived from the CVA scheme and those from the ECMWF model and found that surface precipitation rates were better simulated in the single-column model (SCM) and cloud-resolving model (CRM) when using the CVA-produced forcing data. Xie et al. (2004) also found that the accuracy of the forcing data derived from the numerical weather prediction model was significantly improved after using the observed constraints processed by CVA. Furthermore, the CVA-produced data are typically used for budget analysis, process studies, and driving SCM/CRM models that can be used to evaluate physical parameterizations of climate model (Tang & Zhang, 2015; Xie et al., 2004, 2003; Zhang et al., 2016).

As the CVA method is less sensitive to the input data sources and shows advantages in processing a variety of measurements, it has been used to analyze measurements collected from various field campaigns, such as the Atmospheric Radiation Measurement (ARM) Program (Zhang et al., 2016; Zhang, Lin, et al., 2001), the Tropical Warm Pool-International Cloud Experiment (Xie et al., 2010), Midlatitude Continent Clouds Experiment (Xie et al., 2014), Green Ocean Amazon Experiment (Tang et al., 2016), and TRMM field campaigns (Schumacher et al., 2007; Xie et al., 2010). Xie et al. (2014) and Schumacher et al. (2007) also applied the CVA method to study the heating and drying profiles over tropical regions. Detailed algorithms and numerical realizations can be found in Zhang and Lin (1997) and Zhang, Xie, et al. (2001). The major CVA algorithms are briefly listed as follows.

\[
\frac{\partial \langle \overline{V_h} \rangle}{\partial t} + \langle \nabla \cdot \overline{V_h} \rangle + f \overline{k} \times \nabla (\overline{\phi}) + \nabla (\overline{\rho}) = \frac{d\theta}{dt} 
\]

(1)

\[
\frac{\partial \langle z \rangle}{\partial t} + \langle \nabla \cdot \overline{V_h} \rangle = R_{TOA} - R_s + L_{prec} + SH + L \frac{\partial \langle q_s \rangle}{\partial t} 
\]

(2)

\[
\frac{\partial \langle q_s \rangle}{\partial t} + \langle \nabla \cdot \overline{V_h} \rangle q_s = E_s - P_{prec} - \frac{\partial \langle q_s \rangle}{\partial t} 
\]

(3)

\[
\langle \nabla \cdot \overline{V_h} \rangle = -\frac{1}{g} \frac{dp_s}{dt} 
\]

(4)

The structure of the study is organized as follows: A concise description of the CVA method is given in Section 2. The analysis subregions, periods, and data networks of the TP and SC are also presented in Section 2. Section 3 describes the synoptic conditions during the analysis periods in the two regions. Section 4 shows large-scale dynamic and thermodynamic structures over the TP and SC. Finally, a summary of the results is provided in Section 5.

2. Methodology and Data

2.1. Analysis Method

The CVA method by Zhang and Lin (1997) was developed to calculate large-scale vertical velocity and advective tendencies of heat and moisture from limited-region sounding measurements. The constraint to the sounding data is achieved by making minimum possible adjustments to the initial sounding analysis. These adjustments are comparable to uncertainties in the original measurements (Zhang & Lin, 1997; Zhang, Lin, et al., 2001). After constraining and adjusting the sounding state variables (the zonal wind component u, the meridional wind component v, the water vapor mixing ratio q, and temperature T) with those variables from the surface and TOA measurements, the column-integrated conservation of mass, heat, moisture, and momentum are satisfied. That is, whatever enters into the air column is equal to whatever exits from the column plus the local storage. Finally, the CVA-produced analysis data are dynamically and thermodynamically consistent. Later Schumacher et al. (2007) and Xie et al. (2010) also applied the CVA method to study the heating and drying profiles over tropical regions.
where \( \langle X \rangle = \frac{1}{g} \int g \langle X \rangle dp \).

Sounding variables on the left-hand side of Equations 1–4 are constrained by the right-hand side variables measured at the surface and TOA. In the above equations, \( V_h \) is the horizontal wind, \( \bar{z}_h \) is the geopotential height, \( \bar{r}_s \) is the surface wind stress, and \( s = c_p T + \delta s \) is the dry static energy. \( P_{net} \) denotes the surface precipitation. \( R_s \) and \( R_{TOA} \) identify the net downward radiative fluxes at the surface and the TOA. \( L \) is the latent heat of vaporization, \( SH \) is the sensible heat flux, \( q_{sl} \) is the cloud liquid water content, \( q \) is the mixing ratio of water vapor, and \( E_r \) and \( p_r \) stand for the evaporation and pressure at the surface, respectively.

The final analyzed state variables (\( u^*, v^*, s^*, q^* \)) are obtained by minimizing the following cost function, which is under the strong constraint of Equations 1–4.

\[
I(t) = \iint_{\Omega} \left[ \alpha_s \left( u^* - u_0 \right)^2 + \alpha_i \left( v^* - v_0 \right)^2 + \alpha_s \left( s^* - s_0 \right)^2 + \alpha_q \left( q^* - q_0 \right)^2 \right] dx dy dp
\]

where superscript \(^*\) denotes analyzed variables, subscript \(^0\) represents initial measurements, and \( \alpha \) is the weighting function associated with the estimation of measurement error covariances. These measuring errors are regarded as the sum of instrument and measurement uncertainties specified by manufacturers and 20% of the temporal standard deviations of the observations (Zhang, Lin, et al., 2001).

Given the small number of observation stations over the analysis region, it is challenging to obtain large-scale fields that should be domain-averaged to represent the regional state. Therefore, a suitable background field is necessary to fill the missing soundings and incorporate limited-region measurements into the analysis grids. The analysis data from ECMWF are reliable and often used as the first guess for data production (Tang et al., 2016; Xie et al., 2006; Zhu et al., 2012). In this study, ERA5 reanalysis data (Hersbach et al., 2020) from ECMWF were used as the background field. The Cressman scheme was utilized to interpolate data between observations and the background and then obtain the initial analysis field from the sounding array (Cressman, 1959). The weighting function of the interpolation relied on the distance \( L \) between an observation station and an analysis grid point as well as the difference between the observation and background. The distance \( L \) used in this study was 50 km. If there was no observation within \( L \) of an analysis grid point, the background field would be taken as the analysis (Xie et al., 2010).

The input data for the CVA scheme can be classified into two groups. One is the adjusted variables, and the other is the constraint variables, which will be shown in Section 2.3. The procedure in the CVA scheme to finish the data analysis consists of four steps (Zhang, Xie, et al., 2001). The first step collects and reorganizes the raw measurements into a standard format for further analysis. The second step includes quality control of the raw data, averaging the data over the analysis domain, filling in missing measurements, and interpolation to the analysis grids and times. The third step adjusts and constrains the sounding state variables and then calculates the large-scale quantities. And the fourth step outputs the analyzed variables. In this study, the domain-averaged output data by CVA include single- and multi-level variables, such as precipitation rate, wind, heat and radiative fluxes, vertical velocity, advective tendencies, and heat and moisture budgets (as shown in Table S1), with a temporal resolution of 1 h and a vertical resolution of 25 hPa. Moreover, because the observed constraints were used in the analysis, the accuracy of the CVA-produced data can be significantly improved and partially depended on the accuracy of the measurements (Xie et al., 2004). Therefore, the analysis data can be treated as an approximate representation of observations to a certain degree (Schumacher et al., 2007; Xie et al., 2003).

### 2.2. Analysis Subregions and Periods

To analyze the large-scale structures and heating profiles over the TP and SC, it is necessary to choose suitable subregions for these two regions. Previous findings pointed out the eastern and southern TP have deep convective systems and frequent precipitation during the summer monsoon season (Fujinami & Yasunari, 2001; Xu, 2013; Qie et al., 2014). Liu et al. (1999) and Uyeda et al. (2001) found that there was usually strong convection at Naqu – which is located in the central part of the TP – during summer. Additionally, the sufficiency of observations needs to be considered. There are sufficient surface observations in SC, es-
especially during summer. However, it is known that the TP lacks observations due to the complex terrain. In order to fill the gap of measurements over the TP, an intensive observational network has been established around Naqu in the TIPEX-III, during which an intensive boundary-layer and sounding measurement campaign was first conducted from July to August 2014 (Zhao et al., 2017, 2018). Hence, an analysis subregion centered at Naqu with a radius of 200 km was selected for the TP (Figure 1a). For SC, Gemmer et al. (2011) and Woo et al. (1997) showed Guangdong Province – which is situated in the east of SC – has active precipitation in summer. Luo et al. (2017) and Zheng et al. (2017) showed that there are three strong rainy centers in Guangdong Province, which are located in the eastern coastal belt zone, the west of the Pearl River Delta, and the Beiijiang River Valley-Pearl River Delta block zone, respectively. Therefore, the study considered the analysis subregion of SC centered at Longmen – one of the strongest rainfall sites in the Beiijiang Riv-
er Valley (Zheng et al., 2017). However, given no surface and sounding measurements over the oceanic regions, the range of the SC subregion was reduced. The SC subregion covered a 150-km-radius circle including most areas of the Beijiang River Valley, some of the eastern coastal area, and the eastern part of the Pearl River Delta (Figure 1a). It has been shown that the size of the analysis domain in the CVA scheme mainly affects the magnitude of the analyzed fields rather than their structures (Waliser et al., 2002; Xie et al., 2014).

Because this study focuses on the comparison of large-scale structures associated with precipitation over the TP and SC, the strength of precipitation is a key factor to be considered. Precipitation increases after the onset of the southwest summer monsoon (Ding & Chan, 2005). Previous studies show that the strong convection and precipitation usually occurs from July to August in the TP and from May to June in SC (e.g., Ding, 1992; Fujinami & Yasunari; 2001; Qie et al., 2014; Xu et al., 2009). Wu et al. (2013) pointed out strong convection over the TP mainly occurs in July and August, especially with a peak frequency in August. Luo et al. (2013) found more active convective and precipitation systems occurring from mid-May to June in SC. Therefore, it is suitable to focus on these months in this study. Besides, since an intensive observational campaign of TIPEX-III was first operated in the TP from July to August 2014 (Zhao et al., 2017, 2018), which could provide ample measurements, the study selected August 2014 as the analysis period in the TP. And the study selected June 2016 as the analysis period in SC, during which an intensive sounding observation operation was also carried out around Longmen. Figure 2 shows there was active precipitation during each analysis month in the TP and SC subregions. For the TP, the precipitation was more frequent throughout the analysis period. For SC, most of the precipitation happened in the early-to-mid June and the last days of June, but the magnitude of precipitation was obviously stronger.

Note that the different analysis periods for the TP and SC can cause uncertainty in the results because of different large-scale background circulation. However, as mentioned above, due to the distinct topographic features and geolocation, the TP and SC are distinct from each other in the circulation and monsoon systems as well as the occurring time and intensity of precipitation (Chen & Bordoni, 2014; Ding, 1992; Qie et al., 2014). Since the primary target of this study is to investigate and compare the large-scale structures during the most active precipitation seasons, the uncertainties caused by different atmospheric backgrounds in different months are hard to avoid. Or it needs more data samples to reduce the uncertainty in further studies. On the other hand, the important intensive measurements during August 2014 in the TP and during June 2016 in SC also allow us to study precipitation features better. With a focus on the months with the most active precipitation and convection, we hope that this study can still reflect the large-scale characteristics between the two regions to a certain extent, despite the uncertainty in the different analysis periods.

In order to distinguish the large-scale characteristics under different rainfall intensities in the two regions, domain-averaged surface rain rates calculated from observations were used to define strong, moderate, and light/no rain categories. The rain rates were divided into average hourly rain rate \( R_1 \) and the maximum hourly rain rate \( R_{1\text{max}} \), where the digit 1 represented 1 h. These two indices were calculated for each precipitation process during the analysis periods. Note that the strength and amount of the surface precipitation in the TP and SC are obviously different, and thus the thresholds of the rain rate for the three rain categories in the two regions should be different. According to the China Meteorological Administration, the strong rainfall satisfies a rate of \( \geq 25.0 \) mm/day. However, in the west inland of China with less rainfall, \( \geq 15 \) mm/day is a more suitable criterion for the strong rainfall. In fact, for those regions with rich rainfall, a rain process with a rate of \( \geq 50.0, 25.0–49.9, 10.0–24.9, \) and \( 0.1–9.9 \) mm/day is defined as a rainstorm, and strong, moderate, and light rainfall, respectively. But for the TP, some studies used \( \geq 25.0, 10.0–24.9, 5.0–9.9, \) and

![Figure 2. Temporal evolutions of domain-averaged observed surface rainfall rate (mm/day) (a) during August 2014 in the TP and (b) during June 2016 in SC. SC, South China; TP, Tibetan Plateau.](image-url)
0.1–4.9 mm/day as the criteria of these four rain categories in summer (e.g., Wang et al., 2011; Zhou et al., 2012). In this study, the rainfall was just divided into three categories (strong, moderate, and light/no rain), so we redefined the thresholds of the rain rate $R_{\text{max}}$ for these rain categories after considering the above conventional criteria (as shown in Table S2). Table 1 lists the thresholds of $R_1$ and $R_{\text{max}}^1$ for each rain category. Note that the thresholds of $R_1$ depended upon each precipitation process that satisfied $R_{\text{max}}^1$. It is found that, in the TP (SC), the strong rain category accounts for 60% (80%) of the overall rainfall throughout the analysis period, the moderate-rain category accounts for 35% (15%), and the light/no rain category represents the rest of the rain distribution.

### 2.3. Data

The input data networks of the TP and SC are presented in Figures 1b and 1c. Because of the need for horizontal spatial interpolation on the analysis points (red dots) situated on the circular boundary, measurements outside the subregions were also necessary. In the TP, the observation network consisted of four sounding stations, 121 surface meteorological stations, 11 boundary-layer stations, and 1 × 1 grid points from the Clouds and the Earth’s Radiant Energy System (CERES) instruments on the Aqua satellite. In SC, the observation network included six sounding stations, 112 surface meteorological stations, and 1 × 1 CERES grid points. The background field consisting of upper-level winds, temperature, and humidity was obtained from the ERA5 reanalysis data, with a spatial resolution of 0.25 × 0.25 and temporal resolution of 1 h.

In the networks, the sounding stations provided upper-level state variables (including winds, temperature, and humidity) to be adjusted and constrained in the CVA scheme. In the TP, the temporal resolution of the sounding measurements was 12 h, while in SC, the temporal resolution was 6 h in early June (before June 15) and 12 h in late June. The surface automatic meteorological stations provided hourly surface measurements including surface precipitation, pressure, winds, temperature, and humidity. Note that in the TP, all 121 surface stations could give the measurement of precipitation, but only 78 stations could give the measurement of other state variables. CERES provided hourly longwave and shortwave radiation fluxes at the surface and the TOA, as well as the column cloud water content. Surface latent heat flux and sensible heat flux are important constraints for the sounding variables. In the TP, the surface heat fluxes were obtained from the boundary-layer stations, with a temporal resolution of 0.5 h. However, in SC, there were no measurements of latent and sensible heat fluxes, thereby they were obtained from ERA5 reanalysis with a temporal resolution of 1 h. These measurements can be divided into two groups, one is sounding measurements to be adjusted, and the other is surface and TOA measurements to constrain the sounding data. Table 2 lists the groups of variables and their data sources.

### Table 1
Rain-Rate Thresholds and Rain Fractions for Strong, Moderate, and Light/No Rainfall in the TP and SC

| Rainfall intensity | $R_1$ (mm/day) | $R_{\text{max}}^1$ (mm/day) | Rain fraction | $R_1$ (mm/day) | $R_{\text{max}}^1$ (mm/day) | Rain fraction |
|-------------------|----------------|--------------------------|--------------|----------------|--------------------------|--------------|
| Strong            | >5.0           | >15.0                    | 60%          | >15.0          | >50.0                    | 80%          |
| Moderate          | 2.0–5.0        | 5.0–15.0                 | 35%          | 5.0–15.0       | 15.0–50.0               | 15%          |
| Light/no          | <2.0           | <5.0                     | 5%           | <5.0           | <15.0                   | 5%           |

Abbreviations: SC, South China; TP, Tibetan Plateau.

### Table 2
Required Variables in the CVA Scheme and Their Data Sources

| Variable types     | Variables                        | Data sources               |
|--------------------|----------------------------------|---------------------------|
| Constraint variables| Surface latent and sensible heat fluxes | Boundary-layer stations for the TP ERA5 for SC |
|                    | Surface radiative fluxes         | CERES                     |
|                    | Surface precipitation            | Surface stations          |
|                    | Surface state variables          |                           |
| Adjusted variables  | TOA radiative fluxes             | CERES                     |
| Background fields   | Upper-level state variables      | Sounding stations         |
|                    | Upper-level state variables      | ERA5                      |

Abbreviations: CERES, Clouds and the Earth’s Radiant Energy System; CVA, constrained variational analysis; SC, South China; TOA, top-of-the-atmosphere; TP, Tibetan Plateau.
3. Synoptic Conditions

In this section, the ERA5 reanalysis data and the CVA-produced data were used to examine the synoptic conditions during the analysis periods in the TP and SC. Figure 3 shows the period-averaged moisture flux, geopotential height, wind, and surface temperature derived from ERA5 in each region. Figure 3a shows the moisture flux and wind at 500 hPa for the TP, while Figure 3b shows them at 850 hPa for SC, after considering different altitudes for the two regions (highland for the TP and coast for SC). Circulation conditions for both regions are at 500 hPa. It can be seen that the TP was under the influence of the East Asian trough, accompanied by south and southwest wind flows and noticeable water vapor flux (Figure 3a). The water vapor was mainly carried by the strong South Asian monsoon from the Arabian Sea and the Bay of Bengal (as shown in Figure S1) and then climbed up the TP to support precipitation (Krishnamurti, 1985; Qie et al., 2014; Webster et al., 1998). At the surface, confluence and convergence of cold and warm air were apparent (Figure 3c), which could cause low-level instability. As for SC, situated in front of the large East Asian trough, the southwest monsoon flow was stronger (Figure 3b). The combined effect of the strong evaporation over the Bay of Bengal, the strong transportation of water vapor over the South China Sea (Figure 3b), and the low-level jet (as shown in Figure S1) provided substantial moisture in SC. Figure 3d shows that SC was dominated by strong warm airflow. Therefore, the vigorous low-level warm and moist airflow over SC could lead to plenty of precipitation.
The temporal evolutions of domain-averaged wind, temperature, and water vapor mixing ratio analyzed by CVA in the TP and SC are displayed in Figure 4. Because of the different terrain height over the two regions, the vertical plotting starts at 600 hPa for the TP while 950 hPa for SC. In the TP, the troposphere was characterized by westerlies, with two westerly jet events (>20 m/s) occurring from August 12 to 15 and from 22 to 26, respectively (Figure 4a). The meridional wind was weaker and changeable, with south wind and north wind alternating between the rainfall and no-rainfall days (Figures 2a and 4c). In SC, the troposphere was
dominated by westerlies in the early-to-mid June (Figure 4b), with relatively weaker westerly jets (>15 m/s) from June 5 to 11. After June 19, easterlies gradually occupied the troposphere. Compared to the TP, south-easterlies over SC were more active (Figure 4d). The analyzed winds over SC demonstrated strong southwest monsoon flow in the early-to-mid June and southeast monsoon flow in late June 2016. The warm, moist layer was deeper over SC than over the TP (Figures 4e–4h), again indicating more potential instability and energy for stronger precipitation in SC.

Hourly 0.1 x 0.1 black body temperature (TBB) data from the FY-2E satellite were used to investigate the cloud and precipitation systems over the two regions (as shown in Figures S2 and S3). TBB is a useful benchmark for the estimation of convection (Laing & Fritsch, 1997; Maddox, 1980). The smaller the TBB is, the higher the cloud top is, and the stronger the convection is. The thresholds of −32°C and −52°C of TBB were usually used as criteria to assess the convection intensity (Maddox, 1983; Ni et al., 2017; Yang et al., 2015). During the analysis periods, most of the precipitation in the TP and SC was caused by the strong convective clouds with a TBB value less than −52°C. These convective clouds mainly came from the southwest and were closely associated with the summer monsoon. For the strong and moderate rainfall in both regions, the cloud systems were strong. And the convection appeared stronger in the nighttime over the TP and in the afternoon and evening over SC. For the light/no rainfall, the cloud systems were clearly weaker. It is found that the strong and moderate rainfall was mainly caused by strong cloud systems passing through the analysis regions, while the light/no rainfall was caused by weak cloud systems (TBB > −22°C) or anvil clouds detrained from the strong cloud systems that passed through the vicinity of the analysis regions.

4. Large-Scale Dynamic and Thermodynamic Structures

4.1. Vertical Velocity and Advective Tendencies

This section focuses on analyzing the domain-averaged vertical velocity and the temperature and moisture advection derived by CVA. Figure 5 shows the evolutions of vertical velocity for the TP and SC and separates
the vertical velocity into strong, moderate, and light/no-rain profiles. Obvious upward motion occurred during increasing precipitation periods, and downward motion occurred during decreasing or break precipitation periods (Figures 5a and 5b). In the TP, the magnitude of the upward motion was equal to that of the downward motion. The vertical velocity profiles for the monthly average, strong, and moderate rain in the TP (Figure 5c) featured upward motion, with their maxima all at around 400 hPa. However, the magnitude of the strong-rain profile was almost twice that of the moderate-rain profile. The light/no-rain profile featured downward motion, with a maximum between 300 and 400 hPa, which tended to be symmetrical to the strong-rain profile. Similarly, in SC, the air column during strong and moderate rainfall was also dominated by the upward motion, with a maximum at around 400 hPa (Figure 5d). Additionally, the intensity of the strong-rain profile was also twice that of the moderate-rain profile in SC. However, although the maxima of the upward motion in the TP and SC were similarly located at the same height, the strong- and moderate-rain profiles in SC had a broader large-value region in the troposphere. Furthermore, the upward motion over SC was also twice as strong as that over the TP, which was consistent with stronger and deeper convection over SC. The relatively shallow convection over the TP was limited by the thin air due to high terrain. By comparison, the light/no-rain vertical velocity profile in SC was really weak, showing weak upward motion in the lower troposphere and weak downward motion in the upper troposphere.

Figures 6a and 6b show the temporal evolutions of the analyzed horizontal temperature advection in the two regions. In the TP, the lower troposphere was mainly dominated by the horizontal warm advection, and the upper troposphere was mainly characterized by the horizontal cold advection. While in SC, the entire troposphere showed weak temperature advection. The profiles of the horizontal temperature advection of the three rain categories are displayed in Figures 7a and 7b. In the TP, these profiles showed advective warming below 400 hPa, with a maximum at around 500 hPa (note that the light/no-rain profile showed weaker advective warming below 500 hPa). Advective cooling was present at the upper levels, with a maximum at around 125 hPa. This cooling was possibly due to the cold air detrained by the strong divergence in the high troposphere. Note that the strong-rain profile in the TP showed the strongest horizontal advective warming (cooling) in the troposphere, while the moderate-rain profile showed moderate advective warming.
in the lower troposphere and the weakest advective cooling in the upper troposphere. The light/no-rain profile showed the strongest advective cooling in the middle troposphere (300–500 hPa).

Compared to the TP, the vertical structure of the horizontal temperature advection in SC was very different (Figure 7b). The averaged profile shape over the TP and SC was approximately opposite. The TP had advective warming at the bottom and cooling at the top, whereas SC had the reverse. It should be noted that the magnitudes of these profiles in SC were clearly smaller than those in the TP. In SC, the lower troposphere displayed weak advective cooling and warming that were distributed alternately in the height, whereas the upper troposphere exhibited relatively stronger advective warming. In addition, the strong-rain profile was the weakest in the troposphere among all of the profiles, especially in the upper troposphere, which could also be due to the cold air detrainment related to the high-level divergence. The reason for the difference in the vertical structure of the horizontal temperature advection between the TP and SC remains unclear, which is likely due to the large diversity of terrains. In fact, the strong solar heating and surface sensible

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**Figure 7.** Profiles of domain-averaged (a), (b) horizontal temperature advection (K/day) and (c), (d) vertical temperature advection (K/day) for monthly average (solid line), strong rain (dash line), moderate rain (dash-dot line), and light/no rain (dot line) in the TP (left) and SC (right). The positive (negative) values represent warm (cold) advection. SC, South China; TP, Tibetan Plateau.
heating over TP due to the big highland played an important role in the atmospheric heating and heat transportation (Chen et al., 2015; Qie et al., 2014; Uyeda et al., 2001), thereby leading to stronger horizontal temperature advection in the entire troposphere.

In contrast to the horizontal advection, the vertical temperature advection was more powerful (Figures 6c and 6d). During precipitation periods, the troposphere was characterized by conspicuous vertical warm advection. The shape of the strong-rain vertical temperature advection profiles in the TP and SC was similar (Figures 7c and 7d), both with a maximum at around 250 hPa, although the magnitude was higher in SC. The moderate-rain profile shared high similarity with the monthly average profile in each region, but in the TP, this profile showed a maximum also at around 250 hPa, while in SC, this profile showed a broader maximum between 250 and 400 hPa. For the light/no-rain profile, SC showed weak vertical advective warming in the lower troposphere and weak cooling in the upper troposphere, while the TP showed strong vertical advective cooling in the entire troposphere with a maximum again at around 250 hPa.

Figure 8 compares the moisture advection between the two regions. Compared to the TP, the horizontal and the vertical moisture advection over SC had drastic differences, especially below 400 hPa. During strong and moderate rainfall, the TP was mainly characterized by the horizontal dry advection (Figure 9a), with a peak between 300 and 400 hPa. Moist advection was weak and limited in the boundary layer. During light/no rainfall, the troposphere of the TP featured the horizontal moist advection, with peaks also between 300 and 400 hPa. And some dry advection occurred at 450–550 hPa. The horizontal moisture advection was weak over the TP. By contrast, over SC, the horizontal dry advection and moist advection was relatively stronger (Figure 9b), which had an alternate distribution in the mid-to-low troposphere. Unlike the TP, the strongest advection over SC occurred near the surface. Such behavior was due to more water vapor and stronger low-level convergence for water vapor during precipitation in SC.

The profiles of the vertical moisture advection over the TP (Figure 9c) had the opposite distribution to those of the horizontal moisture advection (Figure 9a). That is, for the strong and moderate rain as well as the
monthly average, the vertical advection profiles showed advective moistening in the troposphere, whereas the horizontal advection profiles mainly showed advective drying. For the light/no rain, the vertical advection profile showed drying, whereas the horizontal advection profile mainly showed moistening. However, these profiles all had peaks located at around 400 hPa, indicating that the layers around 400 hPa were the key passageway for the moisture transport over the TP. In contrast to the moisture advection profiles of the TP that mainly showed a one-peak structure, those of SC showed a multi-peak structure (Figures 9b and 9d). The air column of SC was largely moistened by the vertical advection, except during the light/no-rainfall period (Figure 9d). The strong-rain profile was the strongest and showed major vertical advective moistening between 350 and 850 hPa, with two prominent peaks. The moderate-rain profile showed its maximum between 700 and 850 hPa. The light/no-rain profile was the weakest and only displayed weak advective moistening near the ground.
4.2. Diabatic Heating and Drying

The diabatic heating and drying can relate large-scale circulation to convection as well as suggest cloud types (although this is not done here) and precipitation intensity. The development of convection and precipitation also determines the redistribution of heat and moisture (Hack & Schubert, 1990; Hartmann et al., 1984). Owing to the different topography, circulation, rainfall rate, and radiation, the TP and SC may show significant differences in the atmospheric heating and drying profiles. This section will examine the temporal evolutions and vertical structures of the CVA-produced apparent heat source $Q_1$ and the apparent moisture sink $Q_2$, which were popularized by Yanai et al. (1973). According to Yanai et al. (1973), $Q_1$ and $Q_2$ follow the equations:

$$Q_1 = Q_{rad} + L(C - E) - \frac{\partial s \omega}{\partial p}$$

$$Q_2 = L(C - E) + L \frac{\partial q \omega}{\partial p}$$

where $Q_{rad}$ is radiative heating, $C$ is condensation, and $E$ is evaporation. The overbar represents a horizontal average over the region, and the prime means a deviation from the average value. $Q_1$ and $Q_2$ can be considered as residuals of heat and moisture budgets of large-scale motion. According to Equations 6 and 7, $Q_1$ results from radiative heating, net latent heating, and vertical convergence of the vertical eddy transport of sensible heat, and $Q_2$ is formed by the net condensation of water vapor and vertical divergence of the vertical eddy transport of moisture, which also indicates the contribution of latent heat released from water vapor condensation to $Q_2$ (Yanai et al., 1973).

Figure 10a shows that the analysis period of the TP had diabatic cooling near the surface and diabatic heating in the mid-to-high troposphere. The $Q_1$ profiles of the TP were top-heavy (Figure 10c), among which the strong- and moderate-rain profiles had two heating peaks in the middle troposphere (around 350 hPa) and the high troposphere (around 125 hPa), respectively. The light/no-rain profile showed a lower and weaker heating peak at 450 hPa and a moderate heating peak also at 125 hPa. The middle-troposphere heating peak was slightly higher than the drying peak (Figure 10d) and the upward motion peak (Figure 5c). However, the magnitudes of the middle-troposphere diabatic heating and drying were similar, suggesting that such a heating maximum was largely attributed to the latent heat released by the water vapor condensation and the strong upward motion. At the levels of 100–150 hPa, the $Q_1$ profiles showed the strongest heating (Figure 10c), but $Q_2$ tended to be nearly 0 (Figure 10d), thus the high-troposphere heating peak in the TP had little relationship with the water vapor condensation. In fact, the high-level cloud was frequent over the TP during August 2014, especially during strong precipitation (as shown in Figure S2; Pang, 2018). Therefore, Pang et al. (2019) indicated that this high-troposphere heating peak was largely due to the latent heat released from the condensation of supercooled water – which was transported by the strong ascending airflow in the deep convective systems – into ice crystals that formed high-level clouds over the plateau.

In addition, the diabatic cooling below 450 hPa in the TP was mainly due to the precipitation reducing the ground-atmosphere temperature difference and thereby substantially decreasing the surface heating (Chen & Bordoni, 2014; Wen et al., 2014). The evaporation of rainwater near the ground could also absorb plenty of heat (Kuo & Anthes, 1984; Pang et al., 2019), further promoting low-level cooling.

In SC, the vertical structure of $Q_1$ was different from that in the TP, with major diabatic heating in the middle troposphere (Figure 11a). The $Q_1$ profiles, except for the light/no-rain profile, presented a one-peak structure with a heating maximum between 400 and 500 hPa (Figure 11c). These middle-troposphere heating maxima in SC were much stronger than those in the TP. Previous studies have pointed out that latent heat was the largest contributor to atmospheric heating (e.g., Bhide et al., 1997; Schumacher et al., 2007; Yanai & Tomita, 1998). Below 600 hPa over SC, the moisture sink was the strongest (Figure 11d), whereas the heat source was relatively weaker (Figure 11c), indicating that the latent heat released by the net condensation of water vapor in the low troposphere did not immediately heat the atmosphere at the same height. Note that the vertical location of the diabatic heating maximum in SC was adjacent to the location of...
of the upward motion maximum (Figure 5d), which again suggested that the vertical structure of diabatic heating was highly related to the vertical motion. This result was consistent with Luo and Yanai (1983) and Yanai et al. (1973). Substantial latent heat generated at the lower levels was rapidly transported upwards by the forceful ascending airflow. For the strong rainfall in both regions, the diabatic heating and drying were the strongest.

4.3. Daily and Diurnal Variations

The daily variations of the surface precipitation, \(<Q_1>\) and \(<Q_2>\) over the TP and SC are presented in Figure 12. Note that \(<Q_1>\) and \(<Q_2>\) were vertical integral of the CVA-produced \(Q_1\) and \(Q_2\), respectively. Correlation coefficients between the three variables are given on the top of the figures. The biggest difference between the TP and SC was the correlation coefficient between \(<Q_1>\) and \(<Q_2>\), which was 0.38 in the TP and 0.73 in SC, indicating that the diabatic heating over SC had a closer correlation with the water vapor condensation, but this correlation was weakened over the TP. Chen et al. (2015) have analyzed the correlations between precipitation, \(<Q_1>\) and \(<Q_2>\) over the TP in the rainy summer season. They found that the strong surface sensible heat flux due to the large ground-atmosphere temperature difference could generate dry convection, and thereby reduce the sensitivity of the diabatic heating to the water vapor condensation and precipitation. The low correlation between \(<Q_1>\) and \(<Q_2>\) also implied that radiative heating and surface sensible heating were important contributors to form \(Q_1\) in the TP. Diabatic heating and drying both had a high correlation with precipitation in the two regions, suggesting that the atmospheric diabatic heating and drying could affect the precipitation to a great degree.

Previous studies showed precipitation in the TP usually happened in the night to early morning, while precipitation in SC usually occurred from the afternoon to night (Fujinami et al., 2005; Yu et al., 2007; Zhou et al., 2007). The diurnal cycles of domain-averaged surface rain rate in the two regions are plotted in Figures 13a and 13b. It shows that precipitation in the TP usually extended from late afternoon to early
morning, peaking from 0100 to 0400 local solar time (LST). In SC, most of the precipitation occurred in the early morning (0400–0600 LST) and the afternoon (1500–1700 LST). Corresponding to the development of precipitation, the upward motion in the TP tended to be strong in the nighttime, and the downward motion occurred in the daytime with decreasing precipitation (Figure 13c). In SC, the upward motion had a substantial increase between 250 and 700 hPa in the afternoon but was weaker in the early morning though precipitation at this time was also strong (Figure 13d). In contrast to the TP, there was almost no downward motion over SC during the analysis period.

Figure 13e shows the diurnal cycle of the diabatic heating $Q_1$ in the TP. It shows that the mid-to-high troposphere of the TP was mainly characterized by diabatic heating most of the time, despite weak diabatic cooling in the morning. When precipitation became strong, the diabatic heating also became strong. However, with increasing precipitation from late afternoon to early morning, the low troposphere below 475 hPa was dominated by strong diabatic cooling. Noticeable diabatic heating was also found at these low levels from 0900 to 1600 LST when precipitation became weak. As for $Q_2$ in the TP (Figure 13g), it showed diabatic drying almost throughout the troposphere with strong precipitation and showed moistening with weak precipitation. The drying (moistening) illustrated by $Q_2$ also suggested the contribution of latent heating (cooling) by the water vapor condensation (evaporation) to $Q_1$. At the low levels (below 475 hPa), $Q_1$ displayed “cooling – heating – cooling” in the diurnal cycle, but $Q_2$ displayed “drying (that is equivalent to heating) – moistening (cooling) – drying (heating).” Such a combination between $Q_1$ and $Q_2$ was largely associated with the surface precipitation (Figure 13a) and the upward motion (Figure 13c), among which the surface precipitation could lead to the surface cooling, and the upward motion could rapidly transfer heat from the lower levels to the upper levels.

By comparison, SC was mainly dominated by diabatic heating in the diurnal cycle except 0000–0300 LST (Figure 13f). This was very different from the TP where the diabatic heating and cooling were both noticeable, indicating that precipitation in the TP had a stronger impact on the diabatic heating structure. The maximum heating in SC occurred between 350 and 500 hPa in the afternoon (1200–1700 LST), which was
indicated the vertical convergence was strong during nighttime over the TP but was different. The upper-troposphere positive advection was present in the upper troposphere. This horizontal advection appeared the strongest during strong rainfall. Whereas in SC, the horizontal temperature advection in the troposphere was really weak, with weak cold advection and weak warm advection in alternation in the lower troposphere. Slightly stronger warm advection occurred in the upper troposphere. Among the profiles of the three rain categories in SC, the strong-rain profile was the weakest, especially in the upper troposphere. Causes for these differences in the horizontal temperature advection between the TP and SC remain unclear, but are possibly due to the large diversity of terrains between the two regions. Compared to the strong rainfall, the TP showed conspicuous downward motion while SC showed weak vertical motion in the lower troposphere and stronger cold advection in the upper troposphere. This also suggested that the strongest convective and precipitation activities happened in the afternoon over SC.

The term \( Q_1 - Q_2 \) represents the radiative effect and subgrid-scale transport of moist static energy, which is shown as below.

\[
Q_1 - Q_2 = Q_{\text{rad}} - \frac{\Theta}{\Theta_p} h
\]

where \( h = s + L q \) is the moist static energy. The diurnal cycle of \( Q_1 - Q_2 \) (Figures 13i and 13j) shared high similarity with that of \( Q_1 \) (Figures 13e and 13f). Tang and Zhang (2015) pointed out the radiative heating/cooling was smaller than the subgrid-scale transport of moist static energy during strong precipitation. Therefore, for the strong precipitation times of the day in the two regions, the term \( Q_1 - Q_2 \) mainly represents the vertical transport of moist static energy by turbulence and convection. In the low-to-mid troposphere, both regions mainly showed negative \( Q_1 - Q_2 \) from late afternoon to early morning and positive \( Q_1 - Q_2 \) at the rest of the time. The positive (negative) \( Q_1 - Q_2 \) indicated the vertical convergence (divergence) of moist static energy. In the upper troposphere, the performance of \( Q_1 - Q_2 \) in the two regions was different. The upper-troposphere positive \( Q_1 - Q_2 \) was strong during nighttime over the TP but during daytime over SC. And the major convergence of moist static energy was above 200 hPa over the TP and between 300 and 600 hPa over SC.

5. Summary

This study aims to investigate and compare the large-scale structures of vertical velocity, advective tendencies, and diabatic heating (\( Q_1 \)) and drying (\( Q_2 \)) corresponding to the precipitation over the TP and SC. The analysis was mainly based on the measurements through the CVA method by Zhang and Lin (1997) and conducted in the typical rainy subregions of the two regions during the active monsoon rainy season. Furthermore, in order to further understand the dynamic and thermodynamic structures under different rainfall intensities, the rainfall in these two regions was divided into strong, moderate, and light/no rain. The principal results of the study are summarized as follows.

1. For both regions, the upward motion during strong and moderate rainfall had a maximum at around 400 hPa, but the intensity of the upward motion during strong rainfall was almost twice that of during moderate rainfall. Moreover, the intensity of the upward motion in South China (SC) was also approximately twice that of the Tibetan Plateau (TP), consistent with stronger precipitation. During light/no rainfall, the TP showed conspicuous downward motion while SC showed weak vertical motion throughout the troposphere.

2. Significant differences in the structure and intensity of the temperature advection were noticeable between the TP and SC. The horizontal temperature advection over the TP was more forceful than that over SC. In the TP, the horizontal warm advection dominated the lower troposphere and stronger cold advection was present in the upper troposphere. This horizontal advection appeared the strongest during strong rainfall. Whereas in SC, the horizontal temperature advection in the troposphere was really weak, with weak cold advection and weak warm advection in alternation in the lower troposphere. Slightly stronger warm advection occurred in the upper troposphere. Among the profiles of the three rain categories in SC, the strong-rain profile was the weakest, especially in the upper troposphere.
Figure 13. Diurnal cycles (24 h) of domain-averaged (a), (b) surface rainfall rate, (c), (d) vertical velocity (hPa/h), (e), (f) $Q_1$ (K/day), (g), (h) $Q_2$ (K/day), and (i), (j) $Q_1 - Q_2$ (K/day) for the analysis periods in the TP (left) and SC (right). SC, South China; TP, Tibetan Plateau.
horizontal temperature advection, the vertical temperature advection was much stronger. The profiles of strong- and moderate-rain vertical temperature advection over the TP and SC were similar in shape, with an advective warming maximum at around 200–300 hPa. While the light/no-rain profile of the TP was characterized by strong cold advection in the entire troposphere, and that of SC was characterized by weak warm advection in the lower troposphere and weak cold advection in the upper troposphere.

(3) The transport of moisture was important for rainfall. During strong and moderate rainfall, the TP was mainly dominated by weak horizontal dry advection with a maximum slightly above 400 hPa. During light/no rainfall, the TP was characterized by weak horizontal moist advection with peaks at 300–400 hPa. For the vertical moisture advection, the strong- and moderate-rain profiles of the TP featured moist advection while the light/no-rain profile featured dry advection. However, all of them also showed their maxima at around 400 hPa. The moisture advection over SC was relatively stronger. Compared to the one-peak structure in the moisture advection profiles under different rainfall intensities in the TP, these profiles in SC presented a multi-peak structure.

(4) The $Q_1$ profiles of strong, moderate, and light/no rain categories in the TP were all top-heavy, with two heating peaks at around 125 and 350 hPa (but 450 hPa for light/no-rain profile), respectively. The middle-troposphere heating peak was associated with the upward motion and the latent heat released by the condensation of water vapor into rainwater, while the high-troposphere heating peak was mainly due to the latent heat released by the condensation of supercooled water into ice crystals, which was linked to the deep convection extending up to the high troposphere. Compared to the TP, the $Q_1$ profiles of SC only had a maximum between 400 and 500 hPa, which was also attributed to the strong upward motion and the latent heat from the water vapor condensation. Moreover, the correlation between vertically column-integrated $<Q_1>$ and $<Q_2>$ was much smaller in the TP than in SC, indicating a weak relationship between the diabatic heating and the water vapor condensation over the TP.

(5) During the analysis periods, precipitation over the TP usually happened from late afternoon to early morning, with a peak between 0100 and 0400 LST, while precipitation over SC usually occurred in the early morning and the afternoon, with two peaks between 0400 and 0600 LST and between 1500 and 1700 LST, respectively. Consistent with the development of precipitation, the large-scale vertical velocity and the diabatic heating and drying had a strong diurnal cycle and showed a significant increase when precipitation increased. The largest difference in the diurnal cycle between the two regions was seen in $Q_1$. In the TP, when precipitation became strong, the upper troposphere (above 475 hPa) was mainly dominated by diabatic heating, but the lower troposphere was characterized by diabatic cooling. When precipitation decreased, the troposphere (especially near the surface) was characterized by diabatic heating. In SC, the troposphere was mainly dominated by diabatic heating except during 0000–0300 LST.

The differences and similarities of these large-scale structures were compared and discussed for the TP and SC in this study. However, there are two things to note. First, the analysis was based on one-month measurements in relatively small subregions, so the data sample was small. Second, the analysis periods in different months and years for the two regions had different large-scale background circulation, which might affect the results. Therefore, the conclusions should be treated with caution. However, it is reasonable to believe that, through comparing the large-scale structures over a typical rainy subregion of the TP and SC during their active precipitation periods, the large-scale characteristics as well as their contrasts over the two subsystems of Asian weather and climate can be further understood. Moreover, long-term CVA-analyzed data for these two regions are currently being processed, and we hope to reduce the uncertainty of these results through more data samples in the future. Furthermore, Schumacher et al. (2007) noted that the shape and structure of $Q_1$ and $Q_2$ profiles were related to the cloud types. Therefore, it is worth investigating the sensitivity of diabatic heating to cloud regimes in these two regions in future studies.

Data Availability Statement

The surface and sounding measurement data can be found at http://data.cma.cn/en. The radiative flux data were archived at Clouds and the Earth's Radiant Energy System (CERES; https://doi.org/10.5067/Terra+Aqua/CERES/SYN1deg-1Hour_L3.004A). The TBB data from FY-2E satellite were obtained from
http://satellite.nsmc.org.cn/portal/site/Data/Satellite.aspx. The ERA5 reanalysis data were available at European Center for Medium-Range Weather Forecasts (ECMWF; https://doi.org/10.24381/cds.adbb2d47 and https://doi.org/10.24381/cds.bd0915c6).

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References

Arakawa, A., & Schubert, W. H. (1974). Interaction of a cumulus cloud ensemble with the large-scale environment, Part I. Journal of the Atmospheric Sciences, 31(3), 674–701. https://doi.org/10.1175/1520-0469(1974)031<0674:ioacce>2.0.co;2

Bhide, U. V., Mujamdar, V. R., Ghanekar, S. P., Paul, D. K., Chen, T.-C., & Rao, G. V. (1997). A diagnostic study on heat sources and moisture sinks in the monsoon trough area during active-break phases of the Indian summer monsoon, 1979. Tellus A: Dynamic Meteorology and Oceanography, 49(4), 455–473. https://doi.org/10.1111/tellusa.v49i4.14683

Chang, C.-P. (2004). East Asian monsoon. Hackensack, NJ: World Scientific Publishing Co. Pte. Ltd. https://doi.org/10.1142/5482

Chen, G. T.-J. (1983). Observational aspects of the Mei-Yu phenomenon in Subtropical China. Journal of the Meteorological Society of Japan, 61, 306–312. https://doi.org/10.2151/jmsj1965.61.2_306

Chen, J., & Bordoni, S. (2014). Orographic effects of the Tibetan Plateau on the East Asian summer monsoon: An energetic perspective. Journal of Climate, 27(8), 3052–3072. https://doi.org/10.1175/jcli-d-13-00479.1

Chen, J., Wu, X., Yin, Y., & Xiao, H. (2015). Characteristics of heat sources and clouds over Eastern China and the Tibetan Plateau in boreal summer. Journal of Climate, 28(18), 7279–7296. https://doi.org/10.1175/jcli-d-14-00859.1

Cressman, G. P. (1959). An operational objective analysis system. Monthly Weather Review, 87, 367–374. https://doi.org/10.1175/1520-0499(1959)087<0367:oaos>2.0.co;2

Ding, Y. (1992). Summer monsoon rainfall in China. Journal of the Meteorological Society of Japan, 70, 373–396. https://doi.org/10.2151/jmsj1965.70.1b_373

Ding, Y., & Chan, J. C. (2005). The East Asian summer monsoon: An overview. Meteorology and Atmospheric Physics, 89(1), 117–142

Ding, Y., Wang, Z., & Sun, Y. (2008). Inter-decadal variation of the summer precipitation in East China and its association with decreasing Asian summer monsoon. Part I: Observed evidences. International Journal of Climatology, 28, 1139–1161. https://doi.org/10.1002/joc.1615

Duan, A. M., & Wu, G. X. (2005). Role of the Tibetan Plateau thermal forcing in the summer climate patterns over Subtropical Asia. Climate Dynamics, 24(7), 793–807. https://doi.org/10.1007/s00382-004-0488-8

Flohn, H. (1968). Contribution to a meteorology of the Tibetan Highlands. Atmospheric Science Paper (Vol. 130). Fort Collins, CO: Colorado State University.

Fu, R., Hu, Y., Wright, J. S., Jiang, J. H., Dickinson, R. E., Chen, M., et al. (2006). Short circuit of water vapor and polluted air to the global stratosphere by convective transport over the Tibetan Plateau. Proceedings of the National Academy of Sciences, 103(15), 5664–5669. https://doi.org/10.1073/pnas.0601584103

Fujinami, H., Nomura, S., & Yasunari, T. (2005). Characteristics of diurnal variations in convection and precipitation over the Southern Tibetan Plateau during summer. Sola, 1, 49–52. https://doi.org/10.2151/sola.2005-014

Fujinami, H., & Yasunari, T. (2001). The seasonal and intraseasonal variability of diurnal cloud activity over the Tibetan Plateau. Journal of the Meteorological Society of Japan, 79(6), 1207–1227. https://doi.org/10.2151/jmsj.79.1207

Gemmer, M., Fischer, T., Jiang, T., Su, B., & Liu, L. L. (2011). Trends in precipitation extremes in the Zhujiang River Basin, South China. Journal of Climate, 24(3), 750–761. https://doi.org/10.1175/2010jcli3717.1

Hack, J. J., & Schubert, W. H. (1990). Some dynamical properties of idealized thermally-forced meridional circulations in the tropics. Meteorology and Atmospheric Physics, 41(1–4), 101–117. https://doi.org/10.1007/bf01026813

Hartmann, D. L., Hendon, H. H., & Houze, R. A. (1984). Some implications of mesoscale circulations in tropical cloud clusters for large-scale dynamics and climate. Journal of the Atmospheric Sciences, 41(1), 113–121. https://doi.org/10.1175/1520-0469(1984)041<0113:ioacce>2.0.co;2

Hj, J., Ju, J., Wen, Z., Li, J., & Jin, Q. (2007). A review of recent advances in research on Asian monsoon in China. Advances in Atmospheric Sciences, 24(6), 972–992. https://doi.org/10.1007/s00376-007-9792-2

Hersbach, H., Bell, B., Berrisford, P., Hirahara, S., Horányi, A., Muñoz-Sabater, J., et al. (2020). The ERA5 global reanalysis. Quarterly Journal of the Royal Meteorological Society, 146, 1999–2049. https://doi.org/10.1002/qj.3803

Jiang, X. L. (2016). Constrained objective analysis over Tibetan Plateau: Method and application. M. S. thesis (in Chinese). Beijing: Chinese Academy of Meteorological Sciences.

Johnson, R. H., & Houze, R. A., Jr. (1987). Precipitating cloud systems of the Asian monsoon. In C.-P. Chang, & T. N. Krishnamurti (Eds.), Monsoon meteorology. Oxford monographs on Geology and Geophysics No. 7 (pp. 298–353). New York, NY: Oxford University Press.

Krishnamurti, T. N. (1985). Summer monsoon experiment – A review. Monthly Weather Review, 113(9), 1590–1626. https://doi.org/10.1175/1520-0499(1985)113<1590:smxaaa>2.0.co;2

Laing, A. G., & Michael Fritsch, J. (1997). The global population of mesoscale convective complexes. Quarterly Journal of the Royal Meteorological Society, 123(538), 389–405. https://doi.org/10.1002/qj.4971235805

Lau, K. M. (1992). East Asian summer monsoon rainfall variability and climate teleconnection. Journal of the Meteorological Society of Japan, 70(1B), 211–242. https://doi.org/10.2151/jmsj1965.70.1b_211

Liu, L. P., Chu, R. Z., Song, X. M., Zhou, Y. J., Feng, J. M., Chen, C. P., et al. (1999). Summary and preliminary results of cloud and precipitation observation in Qinghai-Xizang Plateau in GAME-TIBET. Plateau Meteorology, 18(3), 441–450 (in Chinese).

Luo, Y., & Yanai, M. (1983). The large-scale circulation and heat sources over the Tibetan Plateau and surrounding areas during the early summer of 1979. Part I: Precipitation and kinematic analyses. Monthly Weather Review, 111(5), 922–944. https://doi.org/10.1175/1520-0499(1983)111<0922:tlscah>2.0.co;2

Luo, Y., Wang, H., Zhang, R., Qian, W., & Luo, Z. (2013). Comparison of rainfall characteristics and convective properties of monsoon precipitation systems over South China and the Yangtze and Huai River Basin. Journal of Climate, 26(1), 110–132. https://doi.org/10.1175/jcli-d-12-00100.1

Zhang et al.
Yanai, M., Esbensen, S., & 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(4), 611–627. https://doi.org/10.1175/1520-0469(1973)030<0611:dobpot>2.0.co;2

Yanai, M., & Tomita, T. (1998). Seasonal and interannual variability of atmospheric heat sources and moisture sinks as determined from NCEP-NCAR Reanalysis. *Journal of Climate*, 11(3), 463–482. https://doi.org/10.1175/1520-0442(1998)011<0463:sivoa>2.0.co;2

Yang, K., Wu, H., Qin, J., Lin, C., Tang, W., & Chen, Y. (2014). Recent climate changes over the Tibetan Plateau and their impacts on energy and water cycle: A review. *Global and Planetary Change*, 112, 79–91. https://doi.org/10.1016/j.gloplacha.2013.12.001

Yang, X., Pei, J., Huang, X., Cheng, X., Carvalho, I. M. V., & He, H. (2015). Characteristics of mesoscale convective systems over China and its vicinity using geostationary satellite FY2. *Journal of Climate*, 28(12), 4890–4907. https://doi.org/10.1175/jcli-d-14-00491.1

Yu, R., Zhou, T., Xiong, A., Zhu, Y., & Li, J. (2007). Diurnal variations of summer precipitation over contiguous China. *Geophysical Research Letters*, 34(1). https://doi.org/10.1029/2006gl028129

Zhang, M. H., & Lin, J. L. (1997). Constrained variational analysis of sounding data based on column-integrated budgets of mass, heat, moisture, and momentum: Approach and application to ARM measurements. *Journal of the Atmospheric Sciences*, 54(11), 1503–1524. https://doi.org/10.1175/1520-0469(1997)054<1503:cvasdb>2.0.co;2

Zhang, M. H., Lin, J. L., Cederwall, R. T., Yio, J. J., & Xie, S. C. (2001). Objective analysis of ARM IOP data: Method and sensitivity. *Monthly Weather Review*, 129(2), 295–311. https://doi.org/10.1175/1520-0493(2001)129<0295:oaoaid>2.0.co;2

Zhao, P., Li, Y., Guo, X., Xu, X., Liu, L. P., et al. (2017). The third atmospheric scientific experiment for understanding the Earth-atmosphere coupled system over the Tibetan Plateau and its effects. *Arid Land Geography*, 35(1). https://doi.org/10.13826/j.cnki.cn65-1103/x.2012.01.009

Zhou, T., Yu, R., Chen, H., Dai, A., & Pan, Y. (2007). Summer precipitation frequency, intensity, and diurnal cycle over China: A comparison of satellite data with rain gauge observations. *Journal of Climate*, 21(16), 3997–4010. https://doi.org/10.1175/2008JCLI2028.1

Zhu, X., Liu, Y., & Wu, G. (2012). An assessment of summer sensible heat flux on the Tibetan Plateau from eight data sets. *Science China Earth Science*, 55(5), 779–786. https://doi.org/10.1007/s11430-012-4379-2