Characteristics of Heat Sources and Clouds over Eastern China and the Tibetan Plateau in Boreal Summer

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ABSTRACT

In this study, the summer clouds and precipitation over eastern China and the Tibetan Plateau (TP) are examined by analyzing the satellite observations and the apparent heat source $Q_1$ and moisture sink $Q_2$ computed from the NCEP–NCAR reanalysis. The vertically integrated $[Q_1]$ and $[Q_2]$ and precipitation have similar interannual variations in eastern China, revealing the important contribution from the condensation process. This relationship is weakened in east TP (ETP) because of the contribution of the surface sensible heat flux. In west TP (WTP), $[Q_1]$ is negatively correlated with precipitation because the surface sensible heat flux can be sharply weakened by the decrease of ground–air temperature difference due to rainfall. High clouds and deep convection are closely related with $[Q_1]$ and $[Q_2]$ over eastern China and ETP, while middle clouds and nimbostratus are responsible for the condensation over WTP. During the rainy summer, more convective rains and stronger upward motion appear in eastern China. Greater $Q_1$ and $Q_2$ and stronger upward motion present over ETP, while weaker $Q_1$ and upward motion are observed over WTP in the rainy summer when compared to the dry summer. The cloud-water path over eastern China positively correlates with $[Q_1]$ and $[Q_2]$ over ETP. The deep convection over eastern China also positively correlates with the convection over ETP. These correlations suggest that moisture due to the evaporation of cloud water in anvil clouds detrained from the deep convection over ETP can be transported downstream and benefit the development of convection over eastern China.

1. Introduction

Affected by the East Asian monsoon, China experiences a quasi-steady rain belt during this period. Precipitation from May to September accounts for 60%–85% of the annual total precipitation over China. Therefore, the characteristics of precipitation over the East Asian monsoon areas and the effects of monsoon circulation on rainfall have attracted much attention during the last decade (Zhu and Wang 2001; Zhou and Yu 2005; Wang et al. 2007). A stronger summer monsoon is accompanied by more rainfall in north China, while a weaker summer monsoon is followed by heavier rainfall along the Yangtze River valley. During the monsoon period, moisture over the warm tropical/subtropical oceans is transported to the continent by the monsoon circulation. The enhanced summer rainfall in the middle and lower reaches of the Yangtze River (meiyu) is thought to be due to the plentiful supply of
moisture from the tropical western Pacific along the southern periphery of the western Pacific subtropical high (Zhang 2001). Affected by the monsoon circulation, more convective clouds and precipitation are trigged during the summer monsoon period, which can affect the local hydrological cycle and radiation budget (Ding and Chan 2005; Ding et al. 2007; He et al. 2007).

The monsoon circulation paves the way for the development and production of clouds and precipitation, which play key roles in controlling the daily weather and local climate. On the other hand, previous studies (e.g., Yanai et al. 1973; Arakawa and Schubert 1974; Sui and Yanai 1986; Randall et al. 1989) found that convection and clouds can influence large-scale circulation through the redistribution of heat, moisture, and momentum. Clouds and precipitation are crucial elements in the hydrological cycle that are important in determining the scale of the major atmospheric circulation patterns (Webster 1994). Clouds and precipitation can modify the thermodynamic structure and stability of the atmosphere and affect the monsoon circulation through latent heat release (e.g., Ogura and Cho 1973; Soong and Ogura 1980; Yanai and Johnson 1993; Wang and Qian 2000; Xu et al. 2013). However, the characteristics of cloud vertical structure and its relationship with the heat sources during the summer monsoon period still need to be systematically studied.

As a strong thermal factor in the midtroposphere, the Tibetan Plateau (TP) has a profound effect on the general circulation of the Northern Hemisphere, especially on the East Asian monsoon (e.g., Hsu et al. 2014; Yang et al. 2014). According to the study by Wu et al. (2012), the strong heat flux over TP drives the atmosphere upward like a pump and plays a key role in the onset of the East Asian monsoon. At the onset of the summer monsoon, the wind direction will shift and the south wind will prevail in East Asia. As a result, plenty of water vapor is carried from the adjacent oceans to the East Asian monsoon region and produces significant impacts on the weather and climate of eastern China (e.g., Zhou and Yu 2005).

It has been shown that TP can have a profound influence on the weather and climate of the downstream region (e.g., Hsu et al. 2014; Zhang et al. 2014). Xu et al. (2004) reported that the moisture transport from TP contributes to the mei-yu in the middle and lower reaches of the Yangtze River (MLYR). Many previous studies (e.g., Tao and Ding 1981; Huang and Zhou 2002; Duan and Wu 2005; Duan et al. 2013) also found that the presence of TP can have impacts on the extreme climate events in eastern China. However, it is still unclear through what mechanism TP affects the climate of eastern China. The orographic effect is suggested by Chen and Bordoni (2014) as the main mechanism, while the thermal effect is considered by others (e.g., Wu et al. 2012; Xu et al. 2013). Further study of the clouds, precipitation, and heat source over TP should help advance the understanding of the physical processes involved.

It has been reported that there is more frequent cloud-free sky nowadays in China, using the ground dataset (Qian et al. 2006) and the satellite and reanalysis dataset (Zong et al. 2013). Meanwhile, the low-cloud cover shows an increasing tendency in spring and summer over the Yangtze River delta (Zhao et al. 2014). Kaiser (2000) indicated that the cloud cover of northeastern China experiences a remarkable decrease and that extreme climate events also vary (Wang et al. 2013). Xia (2010) claimed that low clouds tend to have an inverse tendency with the total cloud cover in China, leading to a declining evolution of sunshine duration. Further study by Xia (2013) indicated that different parts of China experience various changes. Extreme climate events, especially rainstorms, present an increased frequency over southern China and MLYR in recent years (Wang and Zhou 2005; Zhai et al. 2005; Chen et al. 2012). Meanwhile, TP experienced evident climate changes (Yang et al. 2014), and total cloud cover of TP shows a significant decreasing trend of $-0.09\%$ decade$^{-1}$, as reported by You et al. (2014). Duan and Wu (2008) reported that the decelerated wind speed induces a declining trend in the surface sensible heat flux (SH), which may further affect the development of clouds in boreal summer. These changes of clouds and precipitation over TP can also influence the weather and climate in the downstream region.

In this study, the climatological behaviors of clouds and precipitation in eastern China and TP are investigated using the reanalysis data and satellite observations. The analysis of the relationship between heat sources, clouds, and precipitation should shed light on the understanding of mechanisms through which TP affects the weather and climate of eastern China. Data and methodology are introduced and described in section 2. The interannual variations of heat sources, cloud, and precipitation during boreal summer and the vertical structures of apparent heat source and moisture sink are presented in sections 3 and 4, respectively. The impacts of TP on eastern China are discussed in section 5. Finally, a summary and discussion are given in section 6.

### 2. Data and methodology

#### a. Data

The major data used in this study are the NCEP–NCAR reanalysis with a resolution of $2.5^\circ \times 2.5^\circ$, which have been widely used in weather and climate research. The NCEP–NCAR reanalysis is reliable compared to observations and ERA-40, according to the study of Hsu and Li (2011).
Another advantage of the NCEP–NCAR reanalysis is that the data present good continuity by using a frozen assimilation system (Kalnay et al. 1996), which meets the requirement of the long-term interannual study.

The cloud properties from ISCCP D2 data are used in the diagnostic analysis. The total cloud coverage is consistent with the surface observations in the mid-latitude (Hahn et al. 2001) and presents a high quality over TP, especially for high clouds (Li et al. 2006). The ISCCP data (Rossow et al. 1996; Rossow and Schiffer 1999) include the monthly mean cloud-top pressure and the fractional cloud coverage of nine cloud types defined by the cloud-top pressure and optical thickness with the spatial resolution of 280 km $\times$ 280 km (about 2.5° $\times$ 2.5°). The datasets cover the period from June 1983 to December 2009. Note that there is almost no low-cloud coverage over TP, according to the ISCCP definition, because of the high elevation. According to the cloud-top pressure and cloud optical thickness, the clouds are classified into several categories as shown in Fig. 1.

The precipitation data used in this study are from the Global Precipitation Climatology Project (GPCP) (Adler et al. 2003; Huffman et al. 2009). Version 2 has monthly data extended back to January 1979 with a spatial resolution of 2.5° $\times$ 2.5°. The GPCP dataset shows a good agreement with other precipitation datasets over land (Yin et al. 2004).

### b. Diagnostic methods

Following the studies by Yanai et al. (1973), Yanai and Johnson (1993), and Yanai and Tomita (1998), the apparent heat source ($Q_1$) and the apparent moisture sink ($Q_2$) are calculated by the following equations:

$$Q_1 = c_p \left( \frac{p}{p_0} \right)^\kappa \left( \frac{\partial \theta}{\partial t} + \mathbf{V} \cdot \nabla \theta + \omega \frac{\partial \theta}{\partial p} \right),$$

and

$$Q_2 = -L \left( \frac{\partial q}{\partial t} + \mathbf{V} \cdot \nabla q + \omega \frac{\partial q}{\partial p} \right),$$

where $\theta$ is the potential temperature; $q$ the water vapor mixing ratio; $\mathbf{V}$ the horizontal wind vector; $\omega$ the vertical velocity; $p$ the pressure; $p_0 = 1000$ hPa; $\kappa = R/c_p$; and $R$, $c_p$, and $L$ are the gas constant, the specific heat, and the latent heat of condensation, respectively. The term $\omega$ is recalculated from the horizontal divergence by vertically integrating the continuity equation:

$$\frac{1}{c_p R \cos \phi} \left[ \frac{\partial \mathbf{u}}{\partial \lambda} + \frac{\partial \mathbf{V}}{\partial \phi} (\nu \cos \phi) \right] + \frac{\partial \omega}{\partial p} = 0.$$  

(3)

The boundary condition at the surface ($p = p_s$) is given by

$$\omega = \omega_s = -g \rho (\frac{u_s}{R \cos \phi} \frac{\partial h}{\partial \lambda} + \frac{v_s}{R} \frac{\partial h}{\partial \phi}).$$  

(4)

In Eqs. (3) and (4), $u$ and $v$ are the zonal and meridional wind, respectively, and $\lambda$ and $\phi$ the longitude and latitude, respectively. The value $R$ is the mean Earth radius. The subscript $s$ denotes the surface value. The term $\rho$ is the air density and $h$ the terrain height obtained from the NCEP–NCAR global elevation data. The $Q_1$ and $Q_2$ values are sensitive to the accuracy of vertical velocity. Above the tropopause, the convective heat transport vanishes, and it can be set as follows:

$$Q_1 = Q_R,$$  

(5)

where $Q_R$ is the radiative heating rate. Because the magnitude of $Q_R$ is generally small near the tropopause (Dopplik 1972), the adiabatic condition is imposed at the tropopause as follows (Nitta 1977):

$$\omega = \omega_T = -\left( \frac{\partial \theta}{\partial t} + \mathbf{V} \cdot \nabla \theta \right) \left\{ \frac{\partial \theta}{\partial p} \right\}.$$

(6)

The initial estimate of the horizontal divergence, $D_0$, is adjusted by adding

$$D' = \left( \omega_T - \omega_s - \int_{p_T}^{p_s} D_0 \, dp \right) \left( p_s - p_T \right).$$

(7)

Then the adjusted divergence, $D = D_0 + D'$, is used to calculate the vertical velocity at all levels. Because of horizontally varying surface height and tropopause pressure, the numbers of layers may vary from one grid
to the next. Equations (1)–(7) are used to calculate $Q_1$ and $Q_2$ using NCEP–NCAR 6-hourly datasets.

The parameters $Q_1$ and $Q_2$ can be interpreted following Yanai et al. (1973):

$$Q_1 = \frac{\partial \overline{s}}{\partial t} + \nabla \cdot \overline{s} \mathbf{V} + \frac{\partial \overline{s}}{\partial p},$$

$$= Q_R + L(c - e) - \frac{\partial}{\partial p} \overline{\alpha q'/\omega'},$$

and

$$Q_2 = -L \left( \frac{\partial \overline{q}}{\partial t} + \nabla \cdot \overline{q} \mathbf{V} + \frac{\partial \overline{q}}{\partial p} \right),$$

$$= L(c - e) + L \frac{\partial}{\partial p} \overline{q'/\omega'}, \quad (8)$$

where $S$ is the surface sensible heat and $P$ and $E$ the total rainfall and evaporation, respectively. Usually, $[Q_R]$ is small. The values of $[Q_1]$ and $[Q_2]$ are similar to each other when the heating is mainly due to the condensation process. When $[Q_1]$ is significantly large and $[Q_2]$ is negligible, it implies the presence of dry convection.

The monsoon flow from the ocean affects eastern China from the south to the north. Ding and Chan (2005) summarized the climatological dates of the onset of the Asian summer monsoon in different monsoon regions. South China first experiences the monsoon with the Pearl River delta (PRD) entering the flood season in early May. Then, the monsoon moves northward and brings the mei-yu front at the MLYR, resulting in the famous mei-yu rain belt from the first 10 days to the second 10 days of June. The monsoon advances northward to the north plain of China (NPC), causing the rainiest period of NPC in the year. Finally, it reaches its northern edge in northeastern China (NEC). With the advance northward of the East Asian monsoon, the moisture carried by the monsoon flow will be lost, resulting in different effects on different regions. The land cover of TP shows distinct differences between the east and west regions (e.g., Cui and Graf 2009). The east region is mainly covered by cropped land, while the west region is at present mainly covered by scrubland and bare land. The land cover can affect the behavior of clouds and precipitation through modifying the surface heat fluxes in the summer, which are indicated by Eqs. (10) and (11). Previous studies have shown different features of apparent heat source and moisture sink between west TP (WTP) and east TP (ETP) (e.g., Luo and Yanai 1984). Therefore, the TP is divided into the west and east regions. The extents of WTP and ETP together with four areas (PRD, MLYR, NPC, and NEC) of eastern China are shown in Fig. 2. Since the cloud systems, especially convective clouds, are frequent during the warm and humid season, the boreal summer season (June–August (JJA)) is focused on in this study.

3. Long-term variations of $[Q_1]$, $[Q_2]$, precipitation, and cloud cover

Figure 3 shows the evolution of summer (JJA) anomalies for precipitation, $[Q_1]$, and $[Q_2]$ during the period of 1979–2012. The correlation coefficients between $[Q_1]$ and $[Q_2]$ in eastern China are over 0.84, and the correlation coefficients in ETP and WTP are 0.59 and 0.78, respectively, indicating that the precipitation process is an important contributor to $[Q_1]$ and $[Q_2]$. The correlation coefficients between precipitation, $[Q_1]$, and $[Q_2]$ are over 0.60 (with the highest value of 0.75 between $[Q_2]$ and precipitation in MLYR) in eastern China. The correlations are weaker in ETP, with
correlation coefficients of 0.46 and 0.43 between $Q_1$, $Q_2$, and precipitation compared with those in eastern China. The correlations in WTP are negative, with coefficients of $-0.38$ and $-0.30$.

The consistent variations between the vertically integrated heat source, moisture sink, and summer precipitation over eastern China illustrate that the main heating source is the latent heat released by the formation of precipitation and condensation. The monsoon circulation brings the warm and moist air from oceans and makes a favorable condition for the development of clouds and precipitation in summer. Therefore, $[Q_1]$ and $[Q_2]$ demonstrate strong coupling with the precipitation in eastern China (Figs. 3a–d). When it comes to TP, the situation is more complicated because of the mechanical and thermodynamic effects of TP on large-scale circulation (e.g., Boos and Kuang 2010; Wu 1984; Wu and Zhang 1998; Yanai and Li 1994). Heated by solar radiation, the gigantic bare surface will show much higher temperature than the air in boreal summer because of the lower thermal capacity. The dramatic temperature difference between the land surface and atmosphere results in a strong sensible heat flux injecting into the atmosphere from the ground. This process induces the dry convection and reduces the correlation of the precipitation with $[Q_1]$ and $[Q_2]$ in ETP (Fig. 3f). For WTP, the sensible heat flux is a major heat source during the warm season (e.g., Luo and Yanai 1984). The energy drives the air mass upward, and convective systems develop, most of which are dry convection. This results in a greater difference in values between apparent heat source and moisture sink. Precipitation will moisten the soil and increase the thermal capacity of the land.

**Fig. 2.** Topography (m) of the domain and the selected regions: PRD, MLYR, NPC, NEC, WTP, and ETP, which are marked with boxes.

**Fig. 3.** Evolution of the summer anomalies for precipitation (black solid; mm day$^{-1}$), $[Q_1]$ (gray solid; W m$^{-2}$), and $[Q_2]$ (black dotted; W m$^{-2}$) over the selected regions for 1979–2012. The left ordinate of every panel is for $[Q_1]$ and $[Q_2]$, and the right is for the precipitation.
surface, leading to a reduced temperature difference between the ground and atmosphere. Therefore, the main heat source and driving force of vertical movement will be weakened, leading to an inverse relationship between the apparent heat source and precipitation in WTP (Fig. 3e).

The surface heat flux can influence the growth of clouds and precipitation, and this effect is enhanced over TP because of the highland and thinner atmosphere. Zhu et al. (2012) compared the NCEP–NCAR surface sensible heat fluxes with six other datasets, including the reanalyses (such as ERA-40), the products of the data assimilation system, and the fluxes calculated using the China Meteorological Administration (CMA) observational stations. They found that all datasets show similar interannual variations, although large differences are shown in values. ERA-40 has the fluxes closest to those from the CMA stations. Therefore, ERA-Interim data (Dee et al. 2011) are used to study the summer surface latent heat flux (LH) and sensible heat flux of selected regions as shown in Fig. 4. Since LH relates to the phase change of water vapor and SH mostly depends on the temperature of the land surface, strong LH and weak SH are shown in humid regions, such as PRD and MLYR (Figs. 4a,b). The dry land with low specific heat capacity in NPC and NEC (Figs. 4c,d) will be much warmer and have a larger sensible heat flux than humid regions. The values of SH in TP are comparable to those in north China (e.g., NPC and NEC) and can have a more profound effect on the local energy budget because the atmospheric column over the plateau is much thinner than that over eastern China (Luo and Yanai 1984; Cuo et al. 2013). The considerable SH (Figs. 4e,f) provides a necessary driving force for triggering wet or dry convective cloud systems. The SH calculated using the ground–air temperature difference at stations over central and eastern TP shows a declining tendency from 1980 (Wang et al. 2012), which agrees with the results from the ERA-Interim. This declining trend in SH over ETP is mainly caused by the decelerated wind speed, according to the studies of Duan and Wu (2008) and Yang et al. (2011). Over WTP, the humid land surface will result in a large LH. However, SH will be weakened by the reduced ground–air temperature difference because of increasing precipitation. Therefore, an inverse relationship of strong negative LH and positive SH anomalies are obtained in the dry summer of 2009 (Figs. 3e and 4e).
Figure 5 shows the evolution of summer anomalies for cloud cover over different regions. As the surface pressure of TP is always lower than 600 hPa, low clouds of TP are not displayed. In all regions, high clouds tend to have a similar evolution to the middle clouds but a negative correlation with low clouds. For eastern China, middle clouds in PRD and MLYR and high clouds in PRD show slightly increased tendencies. Middle clouds in WTP and high clouds in ETP also present increased tendencies and result in increased total cloud cover, which may be responsible for the decline of the surface radiation budget during summer over TP in the last decade (e.g., Shi and Liang 2013). The correlation coefficients between precipitation and high clouds fall in the range of 0.32–0.85, with the highest and lowest coefficients in PRD and WTP, respectively. The range of correlation coefficients between precipitation and total cloud cover is 0.24–0.72. This is because precipitation depends not only on the cloud amount but also on the cloud thickness. The larger cloud-water content often relates to much deeper and vigorous clouds.

Figure 6 shows the linear regression analysis between $Q_1$, $Q_2$, and cloud cover. The formation of precipitation requires the presence of clouds, and the associated latent heat release and moisture transport will contribute significantly to $Q_1$ and $Q_2$. Therefore, the total cloud cover generally presents positive correlation with $Q_1$ and $Q_2$ in eastern China. The value for $Q_1$ shows a positive correlation with high clouds in PRD, NPC, and NEC. The correlation coefficients are statistically significant, exceeding the 95% level in PRD [Fig. 6a(1)], while the coefficients can be over 0.60 in NPC and NEC, exceeding the 99% level [Figs. 6c(1),d(1)]. The maximum coefficient between $Q_2$ and high clouds can reach 0.74 in NPC [Fig. 6c(2)], and the positive correlations also exceed the 99% level in NEC and ETP [Figs. 6d(2),f(2)]. The rainfall is produced by high clouds in northern China and ETP. Because of the humid environment, the air mass in PRD can reach its lifting condensation level at the lower altitude, causing weaker dependency of $Q_2$ on high clouds. The fluxes $Q_1$ and $Q_2$ present an inverse relationship with low clouds in eastern China, especially in NPC [Figs. 6c(1),c(2)]. This is because precipitation is largely generated from middle or high clouds. More low clouds mean that the atmosphere is relatively stable. Meanwhile, the increase of low clouds will consume the limited moisture.

Contrary to expectations, there is no significant relationship between $Q_1$, $Q_2$, and middle/high/total clouds in MLYR [Figs. 6b(1),b(2)], which suggests that the surface heat flux may make important contributions to $Q_1$ and $Q_2$ under favorable conditions. The summer of MLYR is warm and humid, especially during the mei-yu period. This will lead to strong latent heat being released into the atmosphere. Meanwhile, the contribution of sensible heat to $Q_2$ cannot be ignored, which
disturbs the relationship between the condensation and $[Q_2]$. Two parts of TP show distinct features. The condensation is mainly caused by middle clouds in WTP, while $[Q_2]$ shows a close relationship with high clouds in ETP. However, $[Q_1]$ does not present a similar relationship in ETP, implying that most of the condensation processes relate to high clouds and make great contributions to $[Q_1]$ and $[Q_2]$. The surface heat flux also makes important contributions to $[Q_1]$ in ETP. The surface heat flux is the major force driving the air mass upward in WTP (Luo and Yanai 1984). Therefore, because of the drier environment, there is no sufficient subsequent moisture to promote the development of clouds, preventing high-cloud development.

Fig. 6. Linear regression for $[Q_1]$, $[Q_2]$ (W m$^{-2}$), and TCC (gray cross), LCC (cyan vertical bar), MCC (red horizontal bar), and HCC (magenta cross) (1983–2009). TCC, LCC, MCC, and HCC are the cloud cover of total clouds, low clouds, middle clouds, and high clouds, respectively. The left and right ordinates of every panel are for clouds at different levels and the total clouds, respectively. The numbers in parentheses are the correlation coefficients. The single asterisk and double asterisk mean that the correlation of corresponding cloud is statistically significant exceeding the 95% and 99% levels, respectively.
and resulting in a close relationship between $Q_2$ and middle clouds in WTP, as shown in Fig. 6e(2).

The relationship between $Q_1$, $Q_2$, and clouds at different levels can reveal where the condensation occurs. The ISCCP dataset not only provides the cloud cover but also classifies the clouds according to the cloud-top pressure and thickness (Fig. 1). This provides an opportunity to examine the relationship between $Q_1$, $Q_2$, and different types of clouds. Among all cloud types, TY01 (cumulus), TY07 (altocumulus), TY09 (nimbostratus), and TY15 (deep convection) show significant cloud cover and a close relationship to $Q_1$ and $Q_2$. Figure 7 shows the fitting relationships between $Q_1$, $Q_2$, and different cloud types. TY01 generally shows a negative relationship with $Q_1$ and $Q_2$, and this relationship is much clearer in NEC. TY15 shows a close relationship to $Q_2$ in PRD, MLYR, NPC, and ETP [Figs. 7a(2),b(2),c(2),f(2)], suggesting that TY15 takes responsibility for $Q_2$ and deep convective clouds play an important role in summer precipitation over these regions. However, $Q_1$ and $Q_2$ show significant positive relationships with TY09, indicating that the condensation in WTP is mainly caused by middle clouds [Figs. 7e(1),e(2)].

Using these criteria, the dry and wet summers are separated from the NM summer, and the samples for each category are shown in Table 1. The mean profiles of $Q_1$ and $Q_2$ for each category and each region are presented in Fig. 8.

The similar vertically integrated values of $Q_1$ and $Q_2$ over eastern China (Figs. 4a–d) suggest that the main heating is due to the condensation processes. Therefore, the $Q_1$ and $Q_2$ profiles of NM (similar to the mean condition) present positive maxima at the levels of 400–500 hPa in PRD, MLYR, and NPC and 400–600 hPa in NEC (Figs. 8a–d). However, the relative positions of peak levels of $Q_1$ and $Q_2$ vary in different regions, implying the main rainfall types of normal climate vary with regions. The peak levels of $Q_1$ and $Q_2$ separate from each other in NEC, denoting the highly convective nature of summer precipitation. The minor separation of $Q_1$ and $Q_2$ peak levels in MLYR suggests that both frontal and convective rains are important to the heating, which agrees with the analysis of Luo and Yanai (1984). The peak levels of $Q_1$ and $Q_2$ are similar in PRD and NPC, showing that the heating is mostly the result of the frontal precipitation in these regions.

In the wet condition, the precipitation results in positive values of $Q_1$ and $Q_2$ over eastern China as shown in Figs. 8a–d, verifying the main heating of this region is due to the condensation process. Compared with the dry condition, the peak levels of $Q_1$ and $Q_2$ under the wet condition show more separated features in NEC, indicating the convective nature of the rains. Further, the separated peak levels of $Q_1$ and $Q_2$ in PRD are observed under the wet condition, implying the rains under this situation are also convective in nature. However, the similar peak levels of $Q_1$ and $Q_2$ reveal that the heating under the wet condition in NPC is mostly due to the frontal rains.

In the dry condition, $Q_1$ and $Q_2$ profiles show opposite structures compared to the wet condition, and the values are almost negative. There are predominantly negative values of $Q_1$ over eastern China, and there is a minor positive value near the surface (Figs. 8a–d), which is due to the surface sensible heat flux. There are negative values of $Q_2$ at around 1000 hPa in eastern China, which indicates stronger evaporation and weaker vertical moisture transport than that of other conditions. The enhanced evaporation is due to the dry environment and will reduce the water content of low clouds, which can delay the precipitation time or even stop precipitation altogether, prolonging low-cloud lifetime and resulting in more low clouds under the dry condition in eastern China. This is partly responsible for the negative relationship between low clouds and $Q_2$ as shown in Figs. 6a–d. The similar peak levels of $Q_1$ and $Q_2$ indicate

4. Vertical structures

To further understand the physical processes responsible for the apparent heat source and moisture sink, the vertical profiles of $Q_1$ and $Q_2$ are examined in this section. In convective cloud systems, the vertical upward motion will lift the moist air above the lifting condensation level, causing the condensation of water vapor and the release of latent heat at higher levels. Therefore, the profiles of $Q_1$ and $Q_2$ will show positive peaks at different levels. Otherwise, $Q_1$ and $Q_2$ will be smaller and have similar profiles if the condensation is largely due to stratus clouds.

Convection usually presents stronger rainfall intensity and is triggered more frequently in the wet and warm summer. To understand the characteristics of summer precipitation, the 34-year (JJA, 1979–2012) rainfalls are grouped into the normal, wet, and dry conditions according to the mean precipitation (avg) and the standard deviation (st) of each region:

1) The normal condition (NM): the summer that area-averaged daily mean rainfall is between $\text{avg} - 0.8\text{st}$ and $\text{avg} + 0.8\text{st}$.
2) The dry condition (dry): the summer that area-averaged daily mean rainfall is less than $\text{avg} - 0.8\text{st}$.
3) The wet condition (wet): the summer that area-averaged daily mean rainfall is greater than $\text{avg} + 0.8\text{st}$.
the frontal nature of the rains under the dry condition in eastern China.

The mean summer profiles in Fig. 8e show that considerable positive $Q_1$ and small $Q_2$ are observed over WTP in all conditions, suggesting that the heating is mainly due to the surface sensible heat flux rather than the surface latent heat flux. Therefore, $Q_1$ decreases rapidly with height, which is similar to the situation near
Table 1. The samples of the NM, dry, and wet conditions for statistical analysis in different regions.

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the surface of the Sahara in summer (Yanai and Tomita 1998). Under this condition, precipitation leads to a change of the surface energy balance and causes subsurface soil cooling (Wen et al. 2014), weakening the surface sensible heat. Therefore, the rainy summer comes with smaller $Q_1$ in WTP (Fig. 3e). However, ETP shows very distinct $Q_1$ and $Q_2$ profiles (Fig. 8f), which is similar to eastern China and clearly indicates that the heating over ETP is due to the latent heat release of condensation. The values for $Q_1$ and $Q_2$ reach their maximums at a similar level. Previous studies (e.g., Luo and Yanai 1983) claimed that the precipitation in ETP is mainly from convection. Therefore, the similar peaks of $Q_1$ and $Q_2$ suggest that convection may be weaker than the normal condition, which is suggested by Qie et al. (2014). In the dry condition, there are positive $Q_1$ (maximum of 0.9 K day$^{-1}$) and negative $Q_2$ values near the surface, showing that the heating due to the sensible heat flux is also important in this region. There are negative values of $Q_1$ and $Q_2$ at upper levels (around 200 hPa) over TP, especially in the wet condition, showing the presence of cooling due to the evaporation of anvil clouds detrained from deep convection.

The values $Q_1$ and $Q_2$ are the residual of heat and moisture budgets of resolvable motion and indicate the presence of cumulus convection with the vertical eddy transports of heat and moisture. The heating profile is highly related to the upward vertical motion (Yanai et al. 1973; Luo and Yanai 1983). Figure 9 shows the mean vertical velocity of different conditions for each region. In eastern China, the wet and dry summer have upward and downward motions, respectively, while the normal summers generally have a vertical velocity in between. In the wet condition, peak levels of vertical velocity in different regions are different. In PRD, MLYR, and NEC, the peak levels are around 500 hPa, and NPC shows a peak level around 300 hPa. For these three regions (PRD, MLYR, and NEC), the vertical velocity increases up to a certain level (around 400–500 hPa), which is due to condensation heating, and the peak levels of vertical velocity are similar with those of $Q_1$ and $Q_2$ in these regions. In eastern China, the upward motion dominates the troposphere while the downward motion presents under the dry condition. The velocity profile of the wet summer in NPC shows a similar shape to the time-averaged profile of Yanai et al. (1973). In the dry condition, there is obviously downward motion at the low level, which indicates the cooling due to evaporation and results in negative $Q_1$ and $Q_2$ values in Figs. 8a–d.

Because of the strong surface heat flux, the vertical velocity is always upward and has a maximum near the surface in TP during summer. Two parts of TP show distinct characteristics. Over WTP, the upward motion is stronger in the dry condition because of the stronger sensible heat flux. The thermal driving force is so strong that the vertical velocity around 450 hPa can be over 1.5 hPa h$^{-1}$ in the dry condition. Because of the single heating source of sensible heat flux and the dry environment, the vertical velocity of WTP is dominated by the surface thermal driving force, leading to a much smaller difference among the three conditions in this region. However, upward motion is observed over ETP in the wet condition, which is a consequence of the surface heat flux and the activities of convection. The maximum vertical velocity over TP in the wet condition occupies a smaller depth than eastern China, indicating that the mature cloud is thinner, which is verified by recent observations (e.g., Luo et al. 2011; Qie et al. 2014).

5. Effects of TP on eastern China

The different characteristics of apparent heat source and moisture sink between TP and eastern China have been discussed in sections 3 and 4. Many studies have confirmed that TP can have profound effects on the atmospheric general circulation and the climate of adjacent and remote regions because of its ultrahigh altitude and expansive area (Ye 1981; Zhou et al. 2009). The Tibetan Plateau affects the occurrence of heavy rain and severe thunderstorms in eastern, southern, and northern China (Tao and Ding 1981) and plays an important role in the summer climate patterns over subtropical Asia through its thermal forcing (Duan and Wu 2005). Meanwhile, the spatial distribution of moisture over TP and its adjacent areas changes remarkably at the onset of the monsoon; TP becomes a humid area at the upper troposphere, and water vapor can be transported downstream along with the westerly flow (Luo and Yanai 1983). Xu et al. (2013) found that the anomalies of spring heating over TP correlate with the summer precipitation patterns in eastern China because of the moisture transport from TP to eastern China. This relationship should be reflected in the cloud behaviors over eastern China. ETP is a more humid area and has more reliable data because it has more ground-based observational stations than WTP. Therefore, the
correlation between cloud observations of eastern China and the $Q_1$ and $Q_2$ of ETP is analyzed to examine the climate teleconnection on cloud behaviors.

To investigate the horizontal moisture transport ($qV$) over TP and adjacent areas, the $qV$ of all levels and the spatial distributions of mean summer $qV$ and wind field at different levels are computed, as shown in Fig. 10. At 300 hPa, the moisture is transported from TP to the regions downstream (e.g., MLYR and NPC). The southwest monsoon flow will assist this transport farther northeast and make impacts on NEC and the Korean Peninsula. There is no significant vapor transport between PRD and TP, but the vapor in south TP can be transported to southern China under favorable conditions, causing extreme climate events (Xie et al. 2010). However, the $qV$ at 850 and 700 hPa shown in Fig. 10

**Fig. 8.** Mean summer profiles of $Q_1$ (K day$^{-1}$) and $Q_2$ (K day$^{-1}$) under different conditions over the selected regions for 1979–2012.
clearly reveals the moisture from the South China Sea and makes great contributions to the water vapor field in PRD, which is in accordance with previous studies (e.g., Liao et al. 2007; Zhou et al. 2010). This is an important part of the moisture transport from both Indian summer monsoon regions across the Bay of Bengal and the tropical western Pacific to East Asia in the lower troposphere, which affects the rainfall in PRD, MLYR, and NPC (Zhang et al. 1999; Zhang 2001). Sufficient water vapor is conducive to the development of clouds, especially for deep convection. Therefore, the horizontal moisture transport at the middle and upper troposphere...
from TP to eastern China makes a favorable environment for the development of deep convective clouds in the downstream region, especially for MLYR.

Figure 11a shows the relationship between summer $[Q_1]$ of ETP and cloud-water path over several regions of eastern China. The results show a positive relationship between $[Q_1]$ of ETP and the cloud-water path of eastern China, suggesting that strong $[Q_1]$ in ETP may lead to the deep clouds in eastern China. The clear moisture transport channel between MLYR and TP
(Fig. 10) is responsible for the greatest correlation between these regions. It is reported that convection over TP is very active during the warm season (Yang et al. 1992; Jiang and Fan 2002; Yang et al. 2014), which results in positive values of [Q_1] and [Q_2] in the warm and rainy summer. The large [Q_1] and [Q_2] values over ETP can induce deeper clouds of the downstream as shown in Figs. 11a,b. There is a close relationship between [Q_2] and the cloud-water path of MLYR, suggesting that wet convection over ETP has more effects on MLYR. However, [Q_1] shows a better relationship with the cloud-water path in MLYR and NEC, implying that further effects on MLYR and NEC are the consequences of both strong sensible heat flux and precipitation. Previous studies have shown that TP can contribute significantly to the rainfall over ETP and MLYR through moisture transport (e.g., Tao and Ding 1981; Xu et al. 2002). Deep convection often leads to the heavy rain events in eastern China during the summer. Figure 11c shows the positive correlations in deep convection between ETP and MLYR, NPC, and NEC, indicating that the exuberant deep convection over ETP can support the development of deep convection in MLYR, NPC, and NEC. The anvils clouds detrained from deep convection over ETP will evaporate because of the strong horizontal wind at high levels, which can moisten the air mass at the upper troposphere. This process may significantly contribute to the moisture transport from TP to eastern China.

6. Summary and discussion

Using the NCEP–NCAR reanalysis and ISCCP data, the characteristics of the heating mechanism, clouds, and precipitation in eastern China and TP are diagnosed and investigated in boreal summer. During the evolution of monsoon circulation, the jet stream from warm southern oceans prevails, affects eastern China in summer, and then loses its strength while moving from the south to the north. Benefitting from the warm and humid air mass associated with the summer monsoon circulation, the heating source in eastern China is mainly due to the condensation process, resulting in positive correlations between the precipitation and the vertically integrated apparent heat source and moisture sink. Because of the extremely high and expansive land, the surface sensible heat flux makes great contribution to the heating source, which reduces the correlation between the precipitation and the heat source in ETP. The precipitation presents different relationships with [Q_1] and [Q_2] in WTP because more rainfall will cool the surface soil and lead to weaker sensible heat flux and evaporation, resulting in smaller values of [Q_1] in the rainy summer.

In eastern China, high clouds show a close relationship with [Q_1] and [Q_2], suggesting that summer precipitation is primarily contributed to by deep convection and high clouds. In the Tibetan Plateau, middle clouds and nimbostratus are responsible for the precipitation in WTP, while high clouds and deep convection are producing the precipitation in ETP. In a normal summer, the precipitation in NEC shows a highly convective nature, while the frontal precipitation is common in PRD and NPC. Both frontal and convective rains are important in MLYR. In a wet summer, the precipitation exhibits a more convective nature in MLYR, NEC, and PRD, but the frontal precipitation contributes more to the rainfall in NPC.

The vertical profiles of Q_1 and Q_2 in ETP are similar to those in eastern China, and the heating is mainly due to the condensation process. The convection over ETP is shallower, and the frontal rains are also important. However, WTP shows considerable positive Q_1 and rather small Q_2 values under all conditions, suggesting that the heating is mainly due to the surface sensible heat flux. There are negative Q_1 and Q_2 values at high levels (around 200 hPa) over ETP, showing the presence of cooling due to the evaporation of anvil clouds detrained from deep convection. Chen and Bordoni (2014) demonstrated that the horizontal advection of dry enthalpy from TP contributes to the moisture convergence downstream. The analysis in this study further suggests that the moisture due to the evaporation of anvil clouds detrained from deep convection over ETP is transported downstream and affects the climate and weather in eastern China.

Several previous studies (e.g., Bao and Zhang 2013; You et al. 2010) have compared the quality of different reanalysis datasets for the analysis of thermodynamic and dynamic aspects over eastern China and TP. We have computed the apparent heat source and moisture sink using ERA-Interim data. Similar interannual variations of vertically integrated heat source and moisture sink are found in the ERA-Interim data when compared with those in the NCEP–NCAR reanalysis. The magnitudes of heating source do show the difference between the two datasets. We will conduct further analysis to evaluate the heat and moisture budgets using both NCEP–NCAR and ERA-Interim data in the ongoing numerical simulation of cloud systems with the cloud-resolving model (CRM) (e.g., Grabowski et al. 1996; Wu et al. 1998; Wu et al. 2008). The large-scale forcing obtained from these two datasets is used to drive the CRM to simulate the cloud systems over eastern China and the Tibetan Plateau.
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