Thermal Effects of the Surface Heat Flux on Cloud Systems over the Tibetan Plateau in Boreal Summer

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ABSTRACT

The influence of surface heat fluxes on the generation and development of cloud and precipitation and its relative importance to the large-scale circulation patterns are investigated via cloud-resolving model (CRM) simulations over the Tibetan Plateau (TP) during boreal summer. Over the lowland (e.g., along the middle and lower reaches of the Yangtze River), the dynamical and thermal properties of the atmosphere take more responsibility than the surface heat fluxes for the triggering of heavy rainfall events. However, the surface thermal driving force is a necessary criterion for the triggering of heavy rainfall in the eastern and western TP (ETP and WTP). Strong surface heat fluxes can trigger shallow convections in the TP. Furthermore, moisture that is mainly transported from the southern tropical ocean has a greater influence on the heavy rainfall events of the WTP than those of the ETP. Cloud microphysical processes are substantially less active and heavy rainfall cannot be produced when surface heat fluxes are weakened by half in magnitude over the TP. In addition, surface heating effects are largely responsible for the high occurrence frequency of convection during the afternoon, and the cloud tops of convective systems show a positive relationship with the intensity of surface heat fluxes.

1. Introduction

Surface radiation and energy budgets are critical components of any land surface model that characterizes hydrological, ecological, and biogeochemical processes (e.g., Liang et al. 2010). In addition, land energy budgets can influence the triggering and development of clouds, which can, in turn, affect local energy budgets and the surface heat fluxes via the modulation of solar radiation, longwave radiation, and precipitation. Particularly, the strong surface heat fluxes can benefit the formation and development of clouds during the warm season. This issue has been paid great attention over the Tibetan Plateau (TP) because of its peculiar thermal characteristics and physical links to the East Asian monsoon, which affects approximately one-third of the

world's population (e.g., Duan and Wu 2005; Hsu et al. 2014; Wang and Liang 2008). The TP is an expansive highland with an average elevation of more than 4000 m above sea level and low vegetation coverage. Consequently, strong surface heat fluxes are directly injected into the middle troposphere, influencing the cloud and precipitation climatology over the TP (e.g., Chen et al. 2015; R. Ma et al. 2018). Furthermore, the thermal effects of the large-scale orography of the TP influence the weather and climate downstream (e.g., Li et al. 2014) and are regarded as a crucial factor in the formation of the South Asian summer monsoon (e.g., Wu et al. 2015).

The surface heat fluxes, including sensible and latent heat fluxes, are among the important factors in the occurrence of boundary layer clouds (e.g., Ek and Holtslag 2004). The moistening and heating of shallow cumuli in the boundary layer can provide favorable conditions for the triggering of deep convection (e.g., Gentile et al. 2013; Wu et al. 2009). Furthermore, the air mass near the

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surface can be lifted upward by the strong surface sensible heat over the TP, which is a branch of the meridional circulation between the TP and the Indian Ocean (e.g., [G. Wu et al. 2007, 2012](#)). According to [Xu et al. \(2014\)](#), the vapor from the tropical ocean can be transported northward and rise up to the TP through a conditional instability of the second kind (CISK)-like mechanism. Previous studies (e.g., [Chen et al. 2015](#); [Luo and Yanai 1984](#)) have verified the important influence of surface heat fluxes on clouds and precipitation over the TP. In addition, different patterns of the diurnal cycle in precipitation have been observed between the TP and other regions (e.g., [Bao et al. 2011](#); [Hu et al. 2010](#); [Xu and Zipser 2011](#)), and differences in diabatic heating (e.g., surface heating effects) have been regarded as the initiator of these differences in the precipitation diurnal cycle between the TP and the lowlands (e.g., [Bao et al. 2011](#)).

Previous modeling and observational studies (e.g., [Jabouille et al. 1996](#); [Wu and Guimond 2006](#)) have shown that the enhancement of surface heat fluxes by the precipitating deep convection is a subgrid process for cloud resolving simulation. A recent study demonstrated that the surface fluxes and radiative heating have opposite effects on precipitation in a radiative-convective equilibrium experiment over the tropical ocean (e.g., [Anber et al. 2015](#)). There are substantial differences in surface heat fluxes between the TP and the lowlands (e.g., East China), for example, the TP shows greater sensible heating fluxes during the daytime and has stronger diurnal cycle signal in surface heating fluxes (e.g., [Chen et al. 2015](#)). Ground-based observations at a station atop the TP showed that the surface sensible heat flux dominates the energy transfer from Earth's surface into the atmosphere, which can significantly influence convective clouds (e.g., [Zhou et al. 2011](#)). Furthermore, an accelerating warming trend has appeared over the TP under the background of global warming (e.g., [Duan and Xiao 2015](#)), which has imposed potential influences on land surface processes and surface heat fluxes. The TP is a great heat source in boreal summer, and this is considered as one of the important driving forces for the active convection during the warm season ([Xu et al. 2014](#); [Jiang et al. 2016](#); [Pan et al. 2019](#)). The surface heat fluxes (including sensible heat flux and latent heat flux) make an important contribution to the heating source in boreal summer (e.g., [Chen et al. 2015](#)). The accurate estimation of surface heat fluxes is important for weather and climate simulations over the TP. However, different datasets show different surface heat fluxes values over the TP (e.g., [Zhu et al. 2012](#)) that can affect cloud and precipitation systems. Meanwhile, the responses of the physical processes in

cloud and the relative importance between surface heat fluxes and the large-scale environment continue to require further study for the better understanding of these processes over the TP.

In this study, a series of simulations are conducted to investigate the responses of the cloud system to the surface heat fluxes over the TP. The model and experimental designs are introduced in [section 2](#). The simulation results and interpretation are given in [section 3](#), followed by conclusions and a discussion in [section 4](#).

2. Model descriptions and experimental designs

a. Model descriptions

The model used in this work is a two-dimensional cloud-resolving model (CRM), which is an anelastic cloud model developed by [Clark et al. \(1996\)](#) with imposed large-scale forcing and modifications to physical processes that are important for the long-term simulation of cloud systems (e.g., [Grabowski et al. 1996](#); [X. Wu et al. 1998, 1999, 2007](#)). The Kessler-type bulk warm rain parameterization ([Kessler 1969](#)) and the Koenig–Murray bulk ice parameterization ([Koenig and Murray 1976](#)) are included in the model. Two classes of ice, referred to as ice A and ice B, are considered by the model, for which both the mixing ratio and the number concentration conservation equations are solved. The ice A field (typically associated with unrimed or lightly rimed ice particles) is formed by the heterogeneous nucleation of pristine ice crystals, and ice B is usually associated with heavily rimed particles (e.g., graupel) of fast-falling velocity and high-density, that originate from the interaction of the rain field with ice A. The surface heat fluxes in the model are calculated by the parameterization based on the study of [Liu et al. \(1979\)](#). However, this parameterization was developed for tropical ocean regions and is not necessarily suitable over land. Surface heat fluxes are determined mainly by land surface properties (i.e., soil temperature and moisture) and the surface air properties (i.e., temperature, moisture, and wind). Moreover, land surface properties and surface air properties can affect one another, which results in a great challenge in accurately estimating surface heat fluxes. Typically, a land model can estimate surface heat fluxes (e.g., [Niu et al. 2011](#); [Draper et al. 2018](#)). However, the model used in this study is currently not coupled with a land model. For simplification, the surface heat fluxes are prescribed by the ERA-Interim dataset in the model. The more detailed information regarding this model is available in the related literature (e.g., [Clark et al. 1996](#); [Chen et al. 2017](#)).

FIG. 1. Topography of East China and the extents of the three studied areas.

b. Data and experimental design

The large-scale forcing terms used to drive the model mimic the effects of large-scale flow on the temperature and moisture fields as deduced from large-scale budgets (e.g., Yanai et al. 1973; Grabowski et al. 1996). In this study, the large-scale horizontal and vertical advection of temperature and moisture is computed from the ERA-Interim reanalysis (hereafter ERAINT; Dee et al. 2011). The p velocity is recalculated from the horizontal divergence by vertically integrating the continuity equations because of the result's sensibility to the vertical velocity. Detailed descriptions of the calculation processes are available in Yanai and Tomita (1998) and Chen et al. (2015).

According to the study of Ueda et al. (2003), there is a large heat source in the eastern TP with an amplitude about 2 times those over the western plateau in May, and the heating over the western TP becomes weaker in July compared to May. Additionally, the land over the eastern part is mostly covered with vegetation, while it is primarily barren land in the western part (e.g., Cui and Graf 2009). Therefore, the TP is separated into the eastern TP (ETP) and western TP (WTP) (Fig. 1). To highlight the cloud particular responses to the surface heat fluxes over the TP, the middle and lower reaches of the Yangtze River (MLYR) is selected for comparison. The MLYR is a flatland in East China and is famous in the meteorological field for the mei-yu period, which often covers mid-June to mid-July with substantial year-to-year variations (e.g., Lau et al. 1988; Ding and Chan 2005; Xu et al. 2009; Luo et al. 2013; Wang et al. 2018). A 30-day simulation period, which contains several strong convective activities, is selected for each region. A

TABLE 1. Design of the sensitive experiments.

Cases	Surface heat fluxes	Large-scale forcing
Run00	Yes, normal	Yes
Run01	Yes, normal	No
Run02	Yes, doubled	Yes
Run03	Yes, doubled	No
Run04	Yes, halved	Yes
Run05	Yes, halved	No

control simulation (Run00) and five sensitivity simulations (Run01–Run05) are conducted for each region, as shown in Table 1. As discussed by Zhu et al. (2012), there are nonnegligible differences in the values of surface heat fluxes between different datasets, and the difference can be as great as 2 times. Therefore, doubled and halved surface heat fluxes are used in the sensitive simulation. Considering that shallow clouds can be triggered by strong surface heat fluxes, a finer horizontal resolution is needed to capture these small-scale clouds. Therefore, all simulations are conducted in a domain of 600 km with a grid of 500 m and a time step of 5 s. The cloud responses to surface heating effects are investigated by shutting down the effects of large-scale forcing in Run01, Run03, and Run05. The comparison between the simulations with or without large-scale forcing enabled the exploration of the cloud responses to large-scale circulation patterns.

3. Results

Figure 2 shows the time series of the large-scale forcing for temperature and moisture, the horizontal wind field, and vertical velocity over three regions. A strong upward motion generally indicates the convective activity, which can result in heavy rain during the summer. The west wind prevails from 2 km above ground level (AGL) with a strong jet concentrated at the range of 4–12 km AGL over the TP. Compared to the lowlands of East China, the upward motion is more frequent but weaker. Consequently, the large-scale forcing is concentrated below 6 km AGL with weak intensity over the TP. Over the lowlands (Fig. 2a), the strong large-scale forcing of temperature ($>8 \text{ K day}^{-1}$) and moisture ($>6 \text{ g kg}^{-1} \text{ day}^{-1}$) can reach as high as 12 km AGL, which corresponds to the strong upward motion (Fig. 2b).

Figure 3 shows the observed precipitation and short-wave albedo at the top of atmosphere (TOA) as well as the simulated results by the control simulation (Run00). The observed precipitation is domain-averaged daily results from the TRMM 3B40 dataset (Huffman et al. 2007) and the CN05 dataset, which is constructed based on the interpolation of station observations in China

FIG. 2. Time series of the large-scale forcing for (a),(c),(e) temperature (shaded, K day⁻¹) and moisture (contoured, g kg⁻¹ day⁻¹, contour values at -12, -9, -6, -3, 3, 6, 9, 12 with negative values in dashed lines), and (b),(d),(f) the wind field (green contoured for north-south and red contoured for east-west, m s⁻¹, contour values at -18, -12, -6, 6, 12, 18 with negative values in dashed lines) and the vertical velocity (shaded, hPa h⁻¹). (a),(b) The MLYR; (c),(d) the WTP; (e),(f) the ETP.

(Wu and Gao 2013). The domain-averaged 3-h albedo is calculated from the Clouds and the Earth's Radiant Energy System (CERES) dataset (Wielicki et al. 1996). The TOA albedo can be affected by the cloud cover and the vertically integrated cloud water (including cloud water content and cloud ice content). Hydrometeors are typically abundant in deep clouds, resulting in elevated albedos. There were four heavy precipitation processes on 8, 16, 20, and 23 July over the MLYR as demonstrated by TRMM observations (Fig. 3a), which were reproduced by the control run. These rainfall events were responses to the strong large-scale forcing (Figs. 2a,b). The albedo shows correspondingly dramatic variations during these rainfall events (Fig. 3d), suggesting that the rapidly increasing albedo was primarily caused by deep convection over the MLYR. The results are very similar to the previously 3-km resolution simulation with the same period for the same region (Chen et al. 2017).

To assess the simulated results over the TP, a particular precipitation dataset for the TP is included for comparison. This dataset merges the Ensemble Multi-Satellite Precipitation Dataset using the Dynamic

Bayesian Model Averaging scheme (EMSPD-DBMA) over the Tibetan Plateau (Y. Ma et al. 2018a,b). It is noticed that the EMSPD-DBMA agrees with ground-based observation (CN05) well over the ETP while shows less agreement over the WTP (Figs. 3b,c). The satellite may underestimate the little rainfall events, which is more often over the WTP, is a possible reason for this. Meanwhile, the rare ground stations can affect the accuracy of the ground-based observed results (CN05) over the WTP. As shown in Figs. 3b and 3c, the model can simulate the major characteristics of the rainfall evolution with wet biases and the CRM results are closer to the ground-based observation (CN05). Compared to the MLYR, there is less rainfall and larger averaged albedo over the TP, suggesting the appearance of shallow convections. Moreover, there is no dramatic variation in albedo during the rainfall period over the TP, implying that deep convection is relatively weak in intensity and small in size (e.g., Luo et al. 2011; Qie et al. 2014). Notably, the biases of precipitation and albedo increase over the TP. A decrease in reliability of reanalysis (e.g., Bao and Zhang 2013; Chen et al. 2017) is a possible reason for the

FIG. 3. Time series of the three different observed precipitation (TRMM, CN05, and EMSPD-DBMA) and albedo (CERES), and the simulated results of control run (Run00) over the MLYR, ETP, and WTP. The precipitation data represent domain averaged daily results and the albedo data represent domain averaged 3-h results.

exaggerated biases over the TP, particularly for the WTP (e.g., rare ground stations can affect the performance of the reanalysis dataset over the WTP). There are even differences between the three observed datasets over the WTP, suggesting that there are great uncertainties in the retrieval datasets and simulation over the WTP. As denoted by previous studies (e.g., Grabowski et al. 1996), the periodic boundary condition used in this model will cycle clouds in the model domain, which may be a possible reason for the wet biases. Additionally, the eastward phase propagation of precipitation and cloudiness is found downstream of the eastern TP (Wang et al. 2005; Yu et al. 2007; Zhou et al. 2008; Xu and Zipser 2011). The propagation process of precipitation and cloudiness may cause additional wet biases over the TP when a model employs a periodic boundary condition.

As shown in Fig. 3a, there were four heavy rain events during the simulated period and these events could be reproduced by the model only in the presence of large-scale forcing over the MLYR (Figs. 4b,d). When the surface heat fluxes increase from one-half to double (Run04 to Run02), the differences in the domain-averaged precipitation is less than 0.3 mm day^{-1} for these four rainfall events, which is relatively small compared to the amount of precipitation of these events. When large-scale forcing is absent (Run01 in Fig. 4a, Run03 in Fig. 3c, and Run05 in Fig. 3e), these heavy rainfall events cannot be reproduced whether the surface heating is strengthened or weakened. These responses suggest that the occurrences of such heavy rainfall events over the MLYR are controlled by the large-scale thermal and dynamical environment.

Compared to the control run (Run00 in Fig. 3b), the rainfall events over the ETP (e.g., 5–7 June, 16–19 June; or days 2–4 and days 13–16 in Fig. 4) can be generated when large-scale forcing is present (Run02 in Fig. 4b and Run04 in Fig. 4d). When large-scale forcing is absent, heavy rainfall is substantially suppressed (Run01 in Fig. 4a, Run03 in Fig. 4c, and Run05 in Fig. 4e), similar to the lowlands, which indicates the importance of the large-scale environment in the triggering of precipitable clouds over the ETP. Compared to the control run, the rainfall events (domain-averaged precipitation $> 0.1 \text{ mm h}^{-1}$) could be suppressed by 16.5% when the surface heat fluxes are weakened over the ETP (Fig. 4d), suggesting that surface heat fluxes have a positive effect on the precipitation amount of the rainfall events. As shown in Fig. 3c, there was a rainfall event on 19 July (day 16) over the WTP. This process was only reproduced by Run00 and Run02, both of which have large-scale forcing and a nonreduced surface heat flux. Meanwhile, the increased surface thermal heating does not lead to an increasing tendency in precipitation (Run02) and the reduced surface heat fluxes could suppress the precipitation from the rainfall events by 17.0% (Run04). These responses could be attributed to limited moisture because of the dry environment and emphasize the importance of transported moisture for the heavy precipitation over the WTP.

Figure 5 provides the mean vapor fluxes (between 600 and 100 hPa) for the four boundaries of the WTP during the simulation period. As shown in Fig. 5a, the south vapor flux is significantly increasing during 19 July and the vapor flux of the north boundary is at a low level.

FIG. 4. (a)–(e) Time series of precipitation differences between the sensitive experiments (Run01–05) and control Run00 experiments for the MLYR, ETP, and WTP. The abscissas are the days since the simulation started. The left ordinate is for the MLYR and the right is for the ETP and WTP.

Meanwhile, the vapor from the west side is increasing while the east side is also at a high level because of the prevailing westerly flow. The domain net vapor flux shows a peak around 19 July (Fig. 5b), indicating the importance of the transported vapor for this rainfall event. The net vapor flux from south–north shows a greater value than that of the east–west boundary during this rainfall event (rainfall on 19 July) over the WTP. Xu et al. (2003) has reported the water vapor transport at the western boundary of the Tibetan Plateau and considered this as an important player in the torrential rain in the Yangtze River valley. Furthermore, heavy precipitation

can occur as a result of the large-scale forcing even if surface heating is weakened (e.g., 28 July or day 25 of Run04, Fig. 3c) over the WTP. This is mostly a result of the moistened environment because of the vapor transported by large-scale circulation during this period (Figs. 5a,b). Otherwise, strong surface thermal forcing mostly triggers dry convection without the transported moisture over the WTP.

To further investigate the moisture transport from the south during heavy rainfall events in the WTP, the meridional cross sections for zonal mean (80° – 90° E) specific humidity and north–south wind component at the

FIG. 5. Mean vapor fluxes (600–100 hPa) for the four boundaries of WTP during the simulation period. (a) The vapor fluxes for four boundaries of the WTP and (b) the net vapor fluxes from south–north, west–east, and the entire domain.

FIG. 6. Meridional cross section for zonal mean (80° – 90° E) specific humidity (shaded, g kg^{-1}) and north–south wind component (vector, m s^{-1}). The terrain is masked by the averaged surface pressure in the gray shaded area.

times of 0000, 0600, and 1200 UTC 19 July were obtained from the ERAINT, as shown in Figs. 6a–c, respectively. Figure 6d shows the mean condition for the entire simulated period. The south wind in the low troposphere blows from the tropical ocean to the north and increases from 0000 to 1200 UTC 19 July (Figs. 6a–c). As a result, the water vapor over the southern slope of the TP increases (Fig. 6c), which causes an increase in humidity (Fig. 6c). Meanwhile, the strong south wind is blocked by the terrain. Then, the airflow will be deflected horizontally and segmental air mass could be forced upward due to the block of the expansion and towering topography of the TP. This process, as well as the triggered convection over the southern slope of the TP, can provide water vapor to the WTP as shown in Fig. 6c, resulting in an increase in the water vapor flux (Fig. 5). According to Xu et al. (2014), there is a two-ladder CISK-like mechanism in this water vapor transport. Figures 5 and 6 show that this vapor transport mechanism is a potentially important player in the heavy rainfall process over the WTP.

Figure 7 shows the temperature, moisture, and vertical velocity differences between the sensitive runs (Run01–05) and control run (Run00). When the surface

heat fluxes are weakened, there are cooling and drying effects on temperature and moisture, respectively, over all regions. Considering the relatively cool and dry atmosphere, the cooling and drying effects are more remarkable over the TP than the lowlands, suggesting that temperature and moisture over the TP are more sensitive to the surface heat fluxes. In addition, there is an inhibiting effect in the vertical velocity in the range of 1.5–4 km AGL when the surface heat fluxes are weakened over the TP (Run04 in Figs. 7h,i). The weakening tendency in vertical velocity as well as the decrease in the atmosphere water-holding capacity (temperature decrease in the range of 0–4 km AGL, Run04 in Figs. 7b,c) takes the main responsibility for the decreasing tendency in water vapor mixing ratio in the range of 0–4 km AGL (Run04 in Figs. 7e,f).

When the effect of the large-scale forcing is absent, there is a slight cooling effect over the MLYR and ETP (Run01 in Figs. 7a,b). However, a warming effect occurs in the WTP. Under the condition of no large-scale forcing effects, the vertical velocities of all three regions weaken (Run01 in Figs. 7g–i) and precipitation decreases (Fig. 4a) over all regions. The large-scale temperature forcing is mostly a cooling effect over the

FIG. 7. Difference between the sensitive run (Run01–05) and control run (Run00) in the vertical profiles for temperature (T , K), water mixing ratio (qv , g kg^{-1}), and vertical velocity (w , ms^{-1}) over the MLYR, ETP, and WTP.

three regions (Figs. 2a–c). The large-scale cooling and moistening can induce cloud and precipitation along with releasing latent heat, which leads to a warming effect and cancel out the large-scale cooling. Meanwhile, more hydrometeor or precipitation makes a favorable condition for the evaporation processes, which prefer to induce a cooling tendency and enhance the cooling effects. The apparent effects are the total effects of the both the direct large-scale cooling and the feedbacks, which could be either cooling or warming. When large-scale forcing is absent, there are cooling effects over the MLYR and ETP. This implies that there is a total warm effect when the large-scale forcing is present, which means that the latent heating effect due to the development of the cloud and precipitation is more important than the cooling effect in these regions. Over the WTP, a warm effect occurs in the absence of large-scale forcing, implying a total cooling effect when the large-scale forcing is present. This suggests that latent heating effect, which is a result of the hydrometeor generation in the development of the cloud and precipitation, shows a weak influence in normal condition over the WTP. Another notable characteristic is that Run03 shows the greatest warming effect (Fig. 7c) over the WTP while Run02 shows the greatest warming effects over the MLYR and ETP (Figs. 7a,b). This corresponds to the strong heating effect because of the surface heat fluxes over the WTP and agrees with the diagnostic results (e.g., Chen et al. 2015).

Over the TP, there are remarkable diurnal variations in various meteorological elements (including the surface heat fluxes) (Yanai and Li 1994; R. Ma et al. 2018). The strong surface thermal driving force cannot only induce convective activities but also lead to deep planetary boundary layer (PBL) over the TP (Chen et al. 2013). Meanwhile, the development of the PBL is strongly related to the convective activities in the boundary layer (Yang et al. 2004). Observational results show that the convection over the TP can evolve from dry shallow convection in the morning to wet deep convection in the afternoon (Yang et al. 2004). Here, the PBL is investigated by comparing the PBL height of sensitive runs and the control run, as shown in Fig. 8. Over the lowlands, the PBL is suppressed during both the daytime and nighttime when large-scale forcing is absent (Run01 and Run05 in Figs. 8a,d). The PBL cloud develop deeper over the TP when the large-scale forcing is absent (Run01 in Figs. 8b,c). Meanwhile, enhanced surface heat fluxes (Run02) can promote the development of the boundary layer (deeper than 1 km) during the daytime over the MLYR (Fig. 8a). Over the TP, stronger surface heat fluxes can induce deeper PBL (Run02 in Figs. 8b,c,e,f). This effect is significant during

the daytime (Run02 in Figs. 8b,c) and is more apparent when the large-scale forcing is extremely weak (Run01 and Run03 in Figs. 8b,c) over the TP. When large-scale forcing is present and the surface heat fluxes are weakened (Run04), these three regions show different responses. There is no significant change in the daytime PBL height over the MLYR while the daytime PBL is suppressed over the TP (Run04 in Figs. 8b,c), by as much as $\sim 50\%$. Compared to the control run, similar daytime PBL heights are simulated by sensitive experiments over the MLYR (Run04 in Fig. 8a). The weakened surface heat fluxes show a total inhibiting effect on the PBL over the ETP (Run04 in Figs. 8b,e). However, there are a inhibiting effect during the daytime and a accelerative effect during the nighttime over the WTP (Run04 in Figs. 8c,f). Usually, the PBL reaches its maximum height during the daytime and its minimum height during the nighttime. There is a strong diurnal cycle in the surface heat flux with strong surface heat fluxes around noon-time and negative sensible heat fluxes during the night over the WTP. These characteristics of the surface heat fluxes over the TP have been reported by previous studies (e.g., Ma et al. 2005). The negative sensible heat flux will impose a cooling effect on the atmosphere and restrains the development of the PBL. Halved surface heat fluxes will weaken the cooling effect and inhibiting effect for the PBL development due to the negative sensible heat flux in the night over the WTP. This is one of the possible reasons for deeper nighttime PBL of Run 04 (Fig. 8f). Meanwhile, this may weaken the suppression effects of the negative surface heat fluxes for the cloud development and may result in an increasing tendency in precipitation over the WTP (days 6, 23, and 25 in Fig. 4d).

Figures 9 and 10 show the heating profiles due to the condensation, evaporation, deposition and freezing, sublimation, and fusion of all cloud systems for different experiments over the ETP and WTP, respectively. The absence of the large-scale forcing causes the weakening of the phase transition in the cloud (Run01 in Figs. 9b and 10b). When the large-scale forcing is absent, the phase transition and its maximum altitude show a positive relationship with the strength of the surface heat fluxes (Run01 and Run03 in Figs. 9c and 10c). As shown in Figs. 9d and 10d, intensive surface heating can promote the development of warm cloud processes because of the heating and upward-motion due to increased energy. The weakened surface heat fluxes cause an inhibitory effect on the development of cloud over the TP regardless of whether large-scale forcing is present (Figs. 9e,f and 10e,f). Furthermore, the weakened surface heat fluxes cause a more dramatic contraction in cloud phase change processes compared to a situation

FIG. 8. Comparison of the 3-hourly and domain averaged (a)–(c) daytime and (d)–(f) nighttime planetary boundary layer height (PBL, km) of different sensitive experiments (Run01–05, ordinate) and control run (Run00, abscissa) over the (a),(d) MLYR; (b),(e) ETP; and (c),(f) WTP.

without large-scale forcing, suggesting that surface heating imposes profound effects on the triggering of deep clouds over the TP. These results endorse the characteristics of precipitation in the corresponding experiment shown in Fig. 4d. A notable characteristic over the WTP is that the warm cloud processes of Run03 are comparable to the those of Run00 (Figs. 10a,d). However, the precipitation of Run03 is declining as shown in Fig. 4b, which implies that clouds formed by the strengthened surface thermal driving force are mostly characterized by the shallow convection of weak precipitation capacity.

Previous studies have indicated that there are diurnal cycles in precipitation and convections over the TP and it has been claimed that the influence of the surface heating is mainly responsible for the diurnal cycle (Sato et al. 2008; Zhou et al. 2008). Figure 11 shows the diurnal cycles of the deep convection cloud top frequency for different experiments over the ETP, WTP, and MLYR. Here, the deep convection cell is defined as the cloud cell with a top of approximately 6–8 km above the ground and a base of

approximately 0.5–2 km above the ground over the TP (Chen et al. 2017). There is no clear diurnal cycle in the MLYR (Fig. 11a–0). However, the control experiments (Run00) show that convections occur frequently around noon (1200 LST) with an increasing tendency in the cloud top height over the TP, which is mainly caused by the strengthened surface heating fluxes. Moreover, the convection shows another high-frequency period around 2100 LST except the peak around noon time over the TP (Figs. 11a–0,b–0). These two peaks generally agree with the precipitation observations of a C-band vertically pointing frequency-modulated continuous-wave radar (R. Ma et al. 2018), in which the echo top height was defined as the maximum height reached by the minimum effective radar reflectivity (−10 dBZ). The convective cloud top is mostly 12–16 km AGL over the MLYR, while it is 8–11 km AGL over the TP. According to the observation at Naqu station in the central TP (R. Ma et al. 2018), the highest observation-period averaged echo top height (maximum height reached −10 dBZ) appears during 1700–1800

FIG. 9. Domain and time (30 day) averaged profiles for the heating effects due to the condensation (Cond, red solid), evaporation (Evap, black dotted), deposition and freezing (Dep + Frez, green solid), sublimation (Sub, blue dash), and fusion (Fus, magenta solid) of different experiments over the ETP. (a)–(f) Run00–Run05, respectively.

LST with a median value of 8.25 km AGL and the large reflectivity (35–40 dBZ) in during a precipitation event can extend up to 11 km AGL. The simulation results agree these observations in the cloud height with difference in the peak time, which may be caused by the definition of the cloud top and the statistical samples of cloud types. A deeper troposphere and affluent moisture lead to more development space and motive force for the convections over the MLYR. Over the MLYR, the diurnal

cycle of deep convection becomes clear and shows a peak around noon (1200 LST in Figs. 11a-1,a-3,a-5) when the influence of large-scale forcing is absent or the surface heat fluxes are enhanced. However, there is no conspicuous diurnal cycle in the deep convection in the control run (Run00 in Fig. 11a-0), implying that large-scale forcing actually dominates the diurnal characteristics of the convections over the MLYR.

FIG. 10. As in Fig. 9, but for the WTP.

Without the influence of large-scale forcing, convection is concentrated during the day, and the convective cloud top height shows an increasing tendency along with the strengthening of the solar radiation over the TP (Figs. 11b-1,b-3,b-5,c-1,c-3,c-5). The intensive surface heating force significantly increases the height of the convection (Run01, Run03, and Run05 in Fig. 11) over the TP. However, the total precipitation amounts of these experiments (Run01, Run03, and Run05 in Fig. 4) are similar in the WTP, which suggest

that the intensive surface heating does not lead to an increasing tendency in rainfall although the deep convection can develop to a higher level. This is mainly because of the limited moisture condition over the WTP. Figures 11b-0 and 11c-0 show that there is another convective peak around 2100–2400 LST over the TP. It is found that this peak disappears when the effects of large-scale circulation are absent (Figs. 11b-1,b-3,b-5,c-1,c-3,c-5), implying that convective clouds during this period are possibly controlled by other

FIG. 11. Diurnal cycle of deep convective cloud top frequency for different experiments over the (a) MLYR, (b) WTP, and (c) ETP.

factors, for example, large-scale circulation patterns and the local circulation (e.g., Fujinami et al. 2005).

4. Conclusions and discussion

The effects of surface heat flux on the development of clouds and precipitation over the TP are discussed via a series of sensitive numerical experiments using the cloud-resolving model. The role of surface heat fluxes in the development of deep convection is investigated and compared to the influences of the large-scale circulation. The highlighted conclusions are as follows.

Large-scale synoptic patterns are mainly responsible for the heavy convective rainfall over East China (e.g., the middle and lower reaches of the Yangtze River) while other factors (e.g., surface heat fluxes) have influence on the triggering of heavy rainfall. However, both large-scale circulation and surface heat flux are the necessary criteria for heavy rainfall over the TP. Intensive surface heating will result in the shallow convective cloud over the TP. However, these clouds are weak in terms of precipitation capacity because of the dry environment. Precipitation could be suppressed by more than 16% when the surface heat fluxes are halved over the TP. The heavy rainfall events over the WTP show a greater dependence on horizontal advection of moisture from outside than over the ETP.

Strong surface thermal forcing has positive effects on the development of PBL and convection over the TP, and it will affect the activeness of the in-cloud microphysical processes and elevates the peak altitude of the warm phase transition processes (e.g., condensation and evaporation). The enhanced surface heat flux can induce a deeper PBL and be benefit to trigger convection, but make little contribution to the precipitation due to the dry environment over the TP. Both the warm and ice cloud microphysical processes (e.g., condensation, sublimation, and deposition) are dramatically restricted when surface heat fluxes are weakened by a half. A diurnal convective cloud top is discovered over the TP. The convection reaches its most energetic period around noon over the TP, which is mainly caused by the diurnal variation in the surface heat fluxes. A convection peak around noon still appears under the condition of weekend surface heat fluxes over the WTP, while this peak diminishes over the ETP under a similar situation.

The relationship between the surface thermal forcing, cloud and precipitation processes is a complex interactive process with feedback and readjustment to another. In this study, surface heat fluxes are obtained from the ERA-Interim reanalysis and

were prescribed and changed simultaneously in the model, which does not consider the interactions between the change in surface heat fluxes and clouds. This could be investigated by coupling a land surface model to the simulating system. Additionally, including ground-based observations of the surface heat flux would help evaluate the model performance over the TP.

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