Genesis of Tibetan Plateau Vortex: Roles of Surface Diabatic and Atmospheric Condensational Latent Heating

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ABSTRACT

Numerical simulations of a nighttime-generated Tibetan Plateau vortex (TPV) are conducted using the advanced Weather Research and Forecasting (WRF) Model. It is found that the nighttime TPV forms as a result of the merging of convections. Although the WRF Model can reproduce the genesis of the nighttime TPV well, colder and drier biases in the lower atmosphere and drier biases in the upper atmosphere are still presented, thus degrading the simulation performance. Intercomparisons among the experiments indicate that the simulations are more sensitive to land surface schemes than to cloud microphysics schemes. The development of convection is more favorable when daytime surface diabatic heating is vigorous. Surface diabatic heating during daytime plays a dominant role in the development of daytime convection and the genesis of nighttime TPV. Further diagnosis of the PV budget reveals that the obvious increase in PV in the lower atmosphere is associated with the evidently strengthened cyclonic vorticity during TPV genesis. This could be attributed to the increased vertical component of net cross-boundary PV fluxes during the merging of convections as well as the significant positive contribution of diabatic heating effects in the lower atmosphere. Therefore, strong daytime surface diabatic heating, which is essential to convection development, could provide a favorable condition for nighttime TPV genesis. Overall results illuminate the complicated process of TPV genesis.

1. Introduction

The Tibetan Plateau vortex (TPV) is an important and frequently observed rain-producing weather system in boreal summer that causes a severe meteorological hazard over the Tibetan Plateau (TP) and its downstream areas. The cyclonic circulation of TPVs is confined primarily to the lower-middle troposphere, with maximum positive vorticity generally present at about 500 hPa. The typical horizontal and vertical scales of TPVs are about 400–500 km and 2–3 km above ground level (AGL), respectively. Most have a warm

core, especially in their early genesis stage, along with a cloudless eye structure dozens of kilometers wide and spiral rainbands that are similar to tropical cyclones. Most of the TPVs originate on the western and central TP (85°–95°N), with a life cycle of several hours to 3 days. Some TPVs complete their whole life cycle with its origins while others move eastward and even move off the plateau to the east (Ye and Gao 1979; Lhasa Group for Tibetan Plateau Meteorology Research 1981; Qiao 1987; Luo et al. 1993).

The physical processes associated with TPV genesis have been investigated from the perspective of both climate (e.g., Sugimoto and Ueno 2010; Feng et al. 2014; Zhang et al. 2014; Li et al. 2014; Liu and Li 2016; Li et al. 2016; Curio et al. 2019) and weather (e.g., Dell’Osso and Chen 1986; Shen et al. 1986a,b; Wang 1987; Wang and Orlanski 1987; Xiang et al. 2013;

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Tian et al. 2015; Yu et al. 2016; Xu and Zhang 2017; Wu et al. 2018; Shou et al. 2019). Most of these previous studies were either based on statistical analysis using datasets from different reanalyses generated by general circulation models (GCMs), synoptic weather maps, and satellite retrievals, or based on numerical simulations using idealized frameworks or coarse model resolutions. It has been recognized that surface diabatic and condensational latent heating, topographical effects of the TP, large-scale circulation such as atmospheric oscillation, the upper-tropospheric jet stream, and so on, all play important roles in the genesis of TPVs. For instance, climatological statistics indicate that TPVs are usually generated in the elevated central and western TP, with significant diurnal variation. The genesis of TPVs is related to the following factors: convergence at 500 hPa associated with northwesterlies from the westerly zone and southerlies from the Bay of Bengal, divergence at 200 hPa associated with the upper westerly jet (Feng et al. 2014; Li et al. 2014), atmospheric oscillation (Zhang et al. 2014), strength of subtropical westerly jet (Curio et al. 2019), atmospheric apparent heating (Liu and Li 2016), and surface diabatic heating (Li et al. 2016). From a weather perspective, Dell'Oso and Chen (1986) showed that condensational latent heating is more important in the development of TPVs, compared to surface sensible heat fluxes; the intensity of TPVs is also influenced by westerly troughs and the southwest monsoon, which is dependent on the terrain height of the TP. Results of Shen et al. (1986a,b) found that surface sensible heating could cause local intensification of vortices over higher elevations; however, these effects were case dependent regarding the position of the upper-tropospheric jet stream before and after the monsoon period. Wang (1987) and Wang and Orlanski (1987) suggested that condensational latent heating is an essential driving force for the development of TPVs; the thermal influence of the elevated plateau topography may appreciably affect vortex initiation by changing the intensity of the forcing associated with the triggering mechanism. Recent numerical simulations (Tian et al. 2015; Xu and Zhang 2017; Wu et al. 2018) with the Weather Research and Forecasting (WRF) Model have been done with the surface diabatic and condensational latent heating artificially turned off for the whole simulation domain and integration. The above results indicate that the relative contributions of surface diabatic and condensational latent heating to TPV genesis are case dependent. Moreover, results based on reanalysis and satellite retrievals suggest that condensational latent heating and convection activity associated with cloud-top fraction and height, cloud phase, and particle size are important during the genesis

and intensification of TPVs (Xiang et al. 2013; Shou et al. 2019).

The above studies consistently highlight the general consensus that TPV genesis could be due to combined dynamic and thermodynamic effects over the TP. Among the factors that influence TPV genesis, however, it is still unclear and commonly argued by researchers how and why surface diabatic and condensational latent heating influence TPV genesis, and which factor is more important. Research on this topic is currently hindered by the lack of detailed and reliable observations, especially in the western TP. Meanwhile, the impacts of surface diabatic and condensational latent heating are case dependent because of the significant diurnal cycle over land, the complex terrain, and the various land surfaces over the TP. Therefore, further investigations are imperative to deepen our understanding of the dynamic and thermodynamic mechanisms of TPV genesis. The purposes of this study are 1) to investigate how and why land surface and cloud microphysics parameterizations influence simulated TPV genesis by way of complex interactions among the physical processes in numerical models and 2) to illustrate the roles and impacts of land surface diabatic and atmospheric condensational latent heating on TPV genesis, especially in a diurnal cycle. Numerical simulations of TPV genesis are performed using the new generation mesoscale community WRF Model. More importantly, we intend to understand the dynamic and thermodynamic mechanisms of TPV genesis.

The next section provides information on the model, data, and experiment design, as well as an overview of TPV case. Section 3 shows model verification against available observations. The dynamic and thermodynamic effects of surface diabatic and condensational latent heating are addressed by diagnosing the potential vorticity (PV) budget, discussed in section 4. Concluding remarks are given in the final section.

2. Descriptions of model, experiment design, data and TPV case

a. Model, experiment design, and data

An advanced research version of the WRF (ARW) Model (Skamarock et al. 2008), version 3.9.1, is employed for numerical simulations. Two domains in a one-way nested procedure are used (Fig. 1a), with horizontal grid spacing of 25 and 5 km. A terrain-following η vertical coordinate is adopted in this study, including 41 vertical levels, with the lowest η level at about 27 m AGL over the TP. The physical parameterization schemes include the Kain–Fritsch cumulus scheme (Kain 2004; for

FIG. 1. Configuration of the (a) two nested WRF simulation domains and (b) terrain height in the innermost domain (d02). The white dots and the text in Fig. 1b denote the location and name of soundings over the TP.

the “d01” domain only); the Rapid Radiative Transfer Model for GCMs (RRTMG; [Iacono et al. 2008](#)) for longwave and shortwave radiation schemes; the Yonsei University (YSU) planetary boundary layer (PBL) scheme ([Hong et al. 2006](#)) and the revised Fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) surface-layer scheme ([Jiménez et al. 2012](#)). Land-cover and land-use data come from the satellite-borne MODIS.

In the WRF Model, surface diabatic heating (surface enthalpy fluxes, or the sum of surface sensible and latent heat fluxes) and condensational latent heating (atmospheric water vapor condensation) are directly related to the parameterizations of land surface and cloud microphysics processes, respectively (these quantities can be derived directly from WRF Model output). Land surface schemes used in this study include Noah ([Chen and Dudhia 2001](#)) and Rapid Update Cycle (RUC; [Benjamin et al. 2004](#)), and cloud microphysics schemes include the WRF single-moment 6-class microphysics scheme (WSM6; [Hong et al. 2004](#)) and Morrison ([Morrison et al. 2009](#)). The simulation with Noah scheme simulates soil variables for four layers with thicknesses of 10, 30, 60, and 100 cm by using the force–restore method, and accounts for detailed physical vegetation and soil hydrology processes and urbanization treatment. The simulation with RUC scheme simulates soil variables for six layers with thicknesses of 0, 5, 20, 40, 160, and 300 cm, and is characterized by a layer method for solving energy and moisture budget equations for soil variables and fluxes. The WSM6 scheme is one moment and prognoses mass mixing ratios for five hydrometeor categories: cloud droplets, rain, cloud ice, snow, and graupel; it assumes

inverse exponential size distributions for rain, snow, and graupel; the ice crystal concentration is diagnosed as a function of cloud ice mass mixing ratio; snow and graupel particles are assumed to be spherical. The Morrison scheme is two-moment and prognoses mass and number mixing ratios of rain, cloud ice, snow, and graupel/hail and mass mixing ratio of cloud droplets; all species except cloud droplets are assumed to follow inverse-exponential size distributions; the cloud droplet size distribution follows a gamma function; ice particles are assumed to be spherical. Detailed comparison and description of these schemes can be found in [Morrison et al. \(2015\)](#) and [Zeng et al. \(2016\)](#). These schemes have been widely used in other studies in simulating weather and climate over the TP (e.g., [Maussion et al. 2011, 2014](#); [Gao et al. 2015](#); [Zeng et al. 2016](#); [Wan et al. 2017](#); [Wang et al. 2019](#)).

Four numerical experiments, NOAH_WSM6, RUC_WSM6, NOAH_MOR, and RUC_MOR (see [Table 1](#)) are conducted on the basis of combining different land surface and cloud microphysics schemes. These experiments are dedicated to 1) comparing the simulation discrepancies in surface diabatic heating obtained by different land surface schemes under the same cloud microphysics scheme, 2) comparing the simulation discrepancies in condensational latent heating obtained by different cloud microphysics schemes under the same land surface scheme, and 3) investigating the impacts of these simulation discrepancies in surface diabatic and condensational latent heating among the experiments on the genesis of TPV. All simulation results will be presented with a 5-km grid to better show vortex evolution.

Initial and boundary conditions are derived from the analyses produced by the National Centers for Environmental

TABLE 1. Configurations of numerical experiments.

Expt name	Land surface schemes	Cloud microphysics schemes
NOAH_WSM6	Noah	WSM6
RUC_WSM6	RUC	WSM6
NOAH_MOR	Noah	Morrison
RUC_MOR	RUC	Morrison

Prediction FNL. The simulation time is from 1200 UTC 27 to 0300 UTC 29 July 2005 (the local time in the study region is UTC + 6h); the first 12 h are regarded as a model spinup period and are not used in the diagnosis. The observations used in this study include conventional atmospheric soundings (Fig. 1b) and cloud-top brightness temperature from the *FY-2C* satellite of the China Meteorological Administration; precipitation data with 0.25° horizontal resolution come from the Tropical Rainfall Measuring Mission (TRMM) Multi-Satellite Precipitation Analysis 3B42 (version 7) products (<https://trmm.gsfc.nasa.gov/>; Zulkafli et al. 2014).

Table 2 shows the pressure levels included in the metadata and compares the terrain height between model and observation for each sounding stations. It indicates that the observed soundings have different elevations, Tuotuo River, Nakqu, and Tingri stations have similar elevations (about 4500 m) and are higher than the Lhasa station (3650 m). The model terrain height is generally higher than the observed terrain height, with the minimum and maximum discrepancies being 21 and 124 m, respectively, suggesting that model terrain elevation is comparable to the real-terrain elevations for these locations. Note that the sunrise and sunset time in the research region during the research period are about 0700–0800 and 2100–2200 local time, respectively.

b. TPV case

Figure 2 shows the whole life cycle of a TPV based on the observed cloud-top brightness temperature from the *FY-2C* satellite. During the daytime, convections over

the TP develop rapidly and become vigorous (Figs. 2a,b). During the nighttime, from 1600 UTC to 2000 UTC 28 July, the TPV generates and strengthens gradually in the central TP around 90°E (Figs. 2c–e); it subsequently decays after 2000 UTC 28 July, with weakened convections (Figs. 2f–h). The TPV completes its life cycle of genesis, intensification, and dissipation with its origins over the TP, and without obvious eastern movement. It is interesting that the convections associated with the nighttime-generated TPV from 1600 UTC to 2000 UTC 28 July are stronger (Figs. 2c–e), with lower cloud-top brightness temperature, relative to other periods, indicating that the convections associated with the nighttime TPV are stronger than the convections during the daytime. In addition, it is also suggested that the genesis of the nighttime TPV near 90°E could be influenced by interactions among the daytime convections, since it probably forms as a result of the aggregation of daytime convections.

3. Simulation results

In this section, the simulations will be compared with the observations in terms of cloud-top brightness temperature, vertical profiles of temperature and humidity, and surface precipitation. Some simulated nonobservable features in relation to TPV genesis will also be discussed.

a. Cloud-top brightness temperature

Figure 2 indicates that the formation of the nighttime TPV and its gradual strengthening from 1600 UTC 28 to 2000 UTC 28 July could be a result of merged convections; it subsequently decays with weakened convections after 2000 UTC 28 July. Therefore, 1200, 1600, and 2000 UTC 28 July are selected to validate model performance in the simulation of TPV genesis.

Figures 3–5 compare the simulated cloud-top brightness temperature among the different numerical experiments at 1200 UTC 28 July, 1600 UTC 28 July, and 2000 UTC 28 July, respectively. Results at 1200 UTC 28 July show that (Fig. 3) daytime convections over the

TABLE 2. Comparison of terrain height (m) of the sounding stations between observation and WRF Model (~900-m horizontal resolution) and the pressure levels (hPa) included in the metadata of each sounding station (also used in generating Fig. 6).

	Terrain height (m)		Pressure levels (hPa)	
	Obs	Model	1200 UTC 28 Jul	0000 UTC 29 Jul
Tuotuo River	4535	4556	586, 500, 400, 300, 250, and 200	587, 500, 400, 300, 250, and 200
Nakqu	4508	4561	588, 500, 498, 400, 331, 300, 250, 218, and 200	588, 500, 400, 300, 250, and 200
Lhasa	3650	3776	650, 500, 400, 300, 250, and 200	653, 500, 400, 300, 250, and 200
Tingri	4302	4426	603, 551, 500, 484, 427, 400, 389, 322, 300, 250, 234, and 200	604, 500, 446, 400, 362, 300, 250, and 200

FIG. 2. Observed cloud-top brightness temperature ($^{\circ}\text{C}$) from the *FY-2C* satellite in relation to the evolution of TPV from 0600 UTC 28 Jul to 0300 UTC 29 Jul 2005. The coastline and borderline of TP are denoted as black contours.

FIG. 3. Simulated cloud-top brightness temperature (shaded; $^{\circ}\text{C}$) and wind vectors at 500 hPa (vectors; m s^{-1}) at 1200 UTC 28 Jul 2005 in (a) NOAH_WSM6, (b) RUC_WSM6, (c) NOAH_MOR, and (d) RUC_MOR. The superposed thin black and blue contours respectively denote the isoheight (geopotential height) of 5830 and 5820 m at 500 hPa. The borderline of TP is denoted as thick black contours.

TP in view of cloud-top brightness temperature are most vigorous in RUC_MOR, followed by NOAH_MOR, RUC_WSM6, and NOAH_WSM6. From 1600 UTC to 2000 UTC 28 July (Figs. 4 and 5) when the nighttime TPV forms and strengthens, apart from similar results in Fig. 3, the most significant discrepancies among the experiments appear in the area of $30^{\circ}\sim 34^{\circ}\text{N}$ and $85^{\circ}\sim 93^{\circ}\text{E}$, where the convections strengthen and assemble with evident decreased cloud-top brightness temperature, especially in RUC_MOR, RUC_WSM6, and NOAH_MOR, consistent with the observations shown in Figs. 2c–e. The simulated cloud-top brightness temperature during the nighttime (Figs. 4 and 5) is lower than that during the daytime (Fig. 3), suggesting strengthened convection intensity during TPV genesis. The observed weakened spiral rainbands of the nighttime TPV during its decay stage are well simulated in RUC_MOR and RUC_WSM6 but are not simulated in the other experiments (not shown). The corresponding comparisons of geopotential height and wind vectors

at 500 hPa among the different experiments show that, both daytime (Fig. 3) and nighttime (Figs. 4 and 5) geopotential heights are lower in the RUC scheme than in the Noah scheme. The closed cyclonic circulation associated with the nighttime TPV forms in the RUC scheme while not shows up in the Noah scheme, and it is stronger with lower geopotential heights in RUC_MOR than in RUC_WSM6 (Fig. 5), indicating that the simulations of daytime convections and the nighttime TPV are sensitive to both land surface and cloud microphysics schemes. However, the impacts of cloud microphysics scheme on TPV simulation depend strongly on the choice of land surface model. Moreover, the simulated wind vectors at 500 hPa over the TP tend to blow toward the central TP, especially during the nighttime in relation to the genesis of the nighttime TPV (Figs. 4 and 5), indicating that the observed merging effects of convections on the genesis of the nighttime TPV could be well reproduced by numerical simulation. Intercomparisons among the experiments indicate that stronger daytime

FIG. 4. As in Fig. 3, but at 1600 UTC 28 Jul 2005.

convections correspond well with the stronger nighttime-generated TPV, implying that the simulation of daytime convections over the TP is essential to the simulation of nighttime TPV genesis, after comparing the evolution of cloud-top temperature between Figs. 2–5 and the geopotential height in Figs. 3–5.

b. Skew T diagram

Figure 6 compares the skew T diagram for the four soundings shown in Fig. 1b at 1200 UTC 28 July (daytime) and 0000 UTC 29 July (nighttime), respectively. Results show that the four experiments have significant simulation biases in both lower and upper atmosphere. Compared to observations, all experiments tend to significantly underestimate temperature and dewpoint in the lower atmosphere below 500 hPa during both daytime and nighttime. Above 500 hPa, although temperature can be well simulated by each experiment, the dewpoint is generally underestimated by all experiments, especially for pressure levels lower than 300 hPa, the simulation discrepancies of dewpoint among different experiments are obvious from 500 to 300 hPa. These results indicate that much colder and drier simulation

biases in the lower atmosphere, and the much drier simulation biases in the upper atmosphere, could be important factors that degrade the simulations of daytime convections and nighttime TPV genesis. Comparisons of the different experiments regarding convective available potential energy (CAPE), pressure of the lifting condensation level P_{LCL} , and temperature at the lifting condensation level T_{LCL} with observations (Table 3) indicate that during the daytime the observed CAPE is not simulated by any experiment; during the nighttime, the observed and simulated P_{LCL} generally increase with increased T_{LCL} and the simulated P_{LCL} and T_{LCL} in all experiments are closer to observations and are better than the simulations during daytime. This is mainly because the simulated temperature and dewpoint profiles, especially in the lower atmosphere below 500 hPa, become closer to observations during nighttime than during daytime. In addition, the simulation discrepancies of P_{LCL} and T_{LCL} among the different experiments are smaller during the nighttime than during the daytime, indicating that the simulation differences are significant during the daytime. Relative to observations, the simulation performances of

FIG. 5. As in Fig. 3, but at 2000 UTC 28 Jul 2005.

different experiments are different at different locations and for daytime and nighttime.

c. Surface precipitation

Figure 7 compares the simulation of the accumulated surface precipitation associated with the nighttime TPV in different experiments against TRMM. Results show that the observed precipitation is located mainly in the central TP, corresponding well with the location of the nighttime TPV. The observed precipitation associated with daytime convections is very weak (not shown), indicating that the nighttime convections associated with the TPV are stronger than the daytime convections and cause evident rainfall. Intercomparisons among the different simulations indicate that the observed rainfall patterns in the central TP are well presented in RUC_MOR and RUC_WSM6, especially in RUC_MOR, but they are not presented in NOAH_MOR and NOAH_WSM6, further indicating the dominant role of the land surface scheme in the simulation of TPV genesis.

To summarize, daytime convections over the TP are essential to the genesis of the nighttime TPV. The simulation

of daytime convections and the nighttime TPV are sensitive to both land surface and cloud microphysics schemes.

4. Effects of surface diabatic and condensational latent heating

As the analysis from the previous sections has shown, the most pronounced simulation discrepancies among the different experiments are the intensity of daytime convections and the genesis of nighttime TPV. Since the land surface and cloud microphysics schemes have direct and fundamental influences on the simulation of surface diabatic and condensational latent heating, in this section, we will compare the simulation discrepancies of surface diabatic and condensational latent heating among the experiments. Then, the respective relationship between the simulated surface diabatic and condensational latent heating and daytime convections and nighttime TPV genesis will be discussed. In particular, the focus will be on the area of 30°–34°N, 85°–93°E because this area generally corresponds well with the location of the nighttime TPV in all experiments.

FIG. 6. Comparison of the skew T diagram for the four soundings (shown in Fig. 1b) for (a),(e) Tuotuo river, (b),(f) Nakqu, (c),(g) Lhasa, and (d),(h) Tingri between the different experiment and observations at (left) 1200 UTC 28 Jul and (right) 0000 UTC 29 Jul 2005.

TABLE 3. Comparisons of CAPE (J), P_{LCL} (hPa), and T_{LCL} (°C) for four soundings between numerical simulations and observations at 1200 UTC 28 Jul and 0000 UTC 29 Jul.

	Tuotuo river			Nakqu			Lhasa			Tingri		
	CAPE	P_{LCL}	T_{LCL}	CAPE	P_{LCL}	T_{LCL}	CAPE	P_{LCL}	T_{LCL}	CAPE	P_{LCL}	T_{LCL}
1200 UTC 28 Jul												
OBS	83.3	456.1	−5.2	4.2	442.8	−5.6	526.5	456.2	−4.2	16.1	468.0	−2.2
NOAH_WSM6	0.0	446.0	−5.8	0.0	469.1	−1.9	0.0	431.9	−6.8	0.0	441.1	−7.0
RUC_WSM6	0.0	463.6	−3.7	0.0	461.7	−2.4	0.0	427.0	−6.9	0.0	445.6	−6.2
NOAH_MOR	0.0	442.4	−6.3	0.0	462.2	−3.2	0.0	414.1	−9.7	0.0	442.9	−6.9
RUC_MOR	0.0	464.5	−3.6	0.0	458.9	−2.6	0.0	413.7	−8.9	0.0	436.5	−7.9
0000 UTC 29 Jul												
OBS	0.0	492.3	−2.2	0.0	477.4	−2.6	0.0	477.3	−3.6	0.0	484.7	−3.4
NOAH_WSM6	0.0	493.0	−1.7	0.0	487.9	−1.9	0.0	490.4	−2.5	0.0	496.2	−2.1
RUC_WSM6	0.0	492.9	−1.8	0.0	498.0	−1.1	0.0	486.4	−3.0	0.0	492.9	−2.6
NOAH_MOR	0.0	489.2	−2.1	0.0	492.8	−2.0	0.0	481.4	−3.5	0.0	489.3	−2.7
RUC_MOR	0.0	499.6	−1.2	0.0	488.5	−2.0	0.0	483.1	−3.7	0.0	479.2	−4.1

a. Area-averaged surface diabatic and condensational latent heating, absolute vorticity, and microphysics particles

Figure 8 compares the time series of area-averaged surface diabatic heating and soil moisture at the first soil layer in the different experiments. Results show that in all experiments the surface diabatic heating is significant during the daytime and is contributed mainly (~70%) by surface sensible heat fluxes (Figs. 8a,b). Surface diabatic heating is much stronger during the daytime than during the nighttime; it is most significant in RUC_WSM6, followed by RUC_MOR, NOAH_WSM6, and NOAH_MOR, indicating that the RUC scheme can produce more significant daytime surface diabatic heating than the Noah scheme. Although the cloud microphysics scheme has an impact on surface diabatic heating, the impact basically depends on the choice of land surface scheme. Figure 8c compares soil moisture at the first soil layer in the different experiments. Results show that the simulation with Noah scheme produces larger soil moisture than the simulation with RUC scheme, especially during the daytime, leading to the weaker surface sensible heat fluxes in Noah than in RUC because of the decreased Bowen ratio, similar to the results of Giorgi et al. (1996) and Eltahir (1998). Therefore, surface diabatic heating is weaker in Noah than in RUC accordingly.

In the WRF Model, surface diabatic heating, which is parameterized by land surface schemes, is usually treated as lower boundary conditions for boundary layer scheme [e.g., see Eq. (B6) of Hong et al. 2006], and thus has a significant and direct influence on convection's development. Figure 9 compares the height–time cross sections of area-averaged θ_e tendency contributed by boundary layer scheme [see Eq. (4) in Hong et al. 2006],

condensational latent heating from cloud microphysics schemes, moist static energy, and absolute vorticity in the different experiments. Results show that significant θ_e tendency contributed by boundary layer scheme and condensational latent heating are located mainly in the middle/upper atmosphere (e.g., 1800–6500 AGL) and lower atmosphere, respectively. The daytime condensational latent heating and moist static energy are most significant in RUC_WSM6, followed by RUC_MOR, NOAH_WSM6, and NOAH_MOR, consistent with the results in Figs. 8a and 8b, indicating that the development of daytime convections is more favorable when surface diabatic heating is vigorous. The simulated absolute vorticity in all experiments is comparable and weaker during daytime, but there is a great discrepancy and strengthening during the nighttime. These features are especially noticeable in the middle and lower atmosphere. In particular, for the same land surface scheme, although surface diabatic and condensational latent heating during the daytime are stronger in the experiments configured with the WSM6 scheme than in those with the Morrison scheme, the mid- and low-level absolute vorticity during nighttime, however, is weaker in the experiments configured with WSM6 scheme than with Morrison scheme. These results could be attributed to the stronger nighttime condensational latent heating in the Morrison scheme than in the WSM6 scheme (also see the related discussion of Fig. 12). Note that the absolute vorticity associated with the nighttime TPV is located in the mid- and low levels and even near the surface; this is because the cyclonic structures of TPV are confined primarily to the lower-middle troposphere, with evident cyclonic convergence flows near the surface, according to Luo et al. (1993).

Figures 3–5 and 9 illustrate that the simulated cloud-top brightness temperature is significantly lower

FIG. 7. Comparison of (a) observed and simulated accumulated surface precipitation in (b) NOAH_WSM6, (c) RUC_WSM6, (d) NOAH_MOR, and (e) RUC_MOR from 1200 UTC 28 Jul to 0300 UTC 29 Jul 2005.

in the Morrison scheme than in the WSM6 scheme. The simulated daytime surface diabatic heating, condensational latent heating and moist static energy, however, are stronger in the WSM6 scheme than in the Morrison scheme. To interpret these simulation discrepancies, the height–time cross sections of the area-averaged mixing ratios of ice (QICE), cloud (QCLOUD), vapor (QVAPOR), snow (QSNOW), and graupel (QGRAUP) in the different experiments are compared in Fig. 10. Results show that the simulations

of microphysical particles are sensitive to both land surface and cloud microphysics schemes. For experiments configured with the same land surface scheme, it is notable that the Morrison scheme tends to produce more QCLOUD than the WSM6 scheme does in the middle and lower atmosphere, especially during the daytime from 0500 UTC 28 July to 1000 UTC 28 July, when surface diabatic heating rapidly increases. This could be the major reason that the simulated surface diabatic heating, condensational latent heating, and

FIG. 8. Comparison of time series of area-averaged (a) surface enthalpy fluxes (W m^{-2}), (b) surface sensible heat fluxes (W m^{-2}), and (c) soil moisture at 1st soil layer ($\text{m}^3 \text{m}^{-3}$) in the different experiments.

moist static energy during the daytime strengthen more evidently in the WSM6 scheme than in the Morrison scheme. In addition, it is also found that the simulated QICE in the upper atmosphere is much higher in the Morrison scheme than in the WSM6 scheme, leading to a much lower simulated cloud-top brightness temperature in the Morrison scheme than in the WSM6 scheme. For experiments configured with the same cloud microphysics scheme, note that compared to Noah, RUC has stronger daytime surface diabatic heating and could result in a significant increase of all microphysical particles. Results prove the dominant role of daytime surface diabatic heating in the intensity of simulated microphysical particles, although the parameterization of the mutual transformation of microphysical particles is very different among the cloud microphysics schemes.

Apart from the analysis above, it is interesting that the simulation discrepancies of surface diabatic and condensational latent heating, moist static energy, microphysical particles, and mid- and low-level vorticity between the different land surface schemes are more significant than those between the different cloud microphysics schemes. For the experiments configured with the same land surface scheme, different cloud microphysics schemes have great discrepancies mainly in the distribution and number of microphysical particles. These results elucidate and highlight the dominant role that surface diabatic heating induced by the land surface scheme plays in the development of daytime convection and the genesis of the nighttime TPV. The contributions from condensational latent heating induced by the cloud microphysics scheme, however, are largely determined by which land surface scheme is configured simultaneously.

b. PV budget analysis in relation to TPV genesis

The above results indicate that the development of the closed cyclonic mesoscale circulation associated with nighttime TPV genesis (e.g., thin blue contours in Fig. 5) is manifested by a significant increase in cyclonic vorticity in the mid- and lower levels (red contours in the lower atmosphere in Fig. 9) and a significant decrease in cloud-top brightness temperature (Figs. 3–5). Although surface diabatic heating during the daytime has been revealed to be essential to the development of daytime convection and the genesis of the nighttime TPV, some very important questions in terms of the nighttime-generated TPV still need to be addressed. For instance, what are the sources of the mid- and low-level closed cyclonic vorticity associated with nighttime generated TPV? Is the vorticity generated locally in relation to diabatic heating effects, or generated from the merging of convections in its vicinity, or related to an upper-level disturbance? How do surface diabatic and condensational latent heating affect its formation? To answer these questions, a PV budget analysis will be presented based on the hourly model output.

Following Koshyk and McFarlane (1996), it can be shown that the potential vorticity PV (PVU; $1 \text{ PVU} = 10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ K kg}^{-1}$) budget equation without the friction effects in the η vertical coordinate can be expressed as

$$\frac{\partial \text{PV}}{\partial t} = -u \frac{\partial \text{PV}}{\partial x} - v \frac{\partial \text{PV}}{\partial y} - \bar{\omega}_\eta \frac{\partial \text{PV}}{\partial \eta} + \frac{g}{\mu} \left[\left(\frac{\partial v}{\partial \eta} \frac{\partial H}{\partial x} - \frac{\partial u}{\partial \eta} \frac{\partial H}{\partial y} \right) - (\zeta_\eta + f) \frac{\partial H}{\partial \eta} \right], \quad (1)$$

FIG. 9. Height–time of area-averaged θ_e tendency contributed by the boundary layer scheme (shaded; K h^{-1}), condensational latent heating (black contours, at intervals of 0.4, 0.8 and 1.2, with negative values omitted; K h^{-1}), absolute vorticity (red contours, at intervals of 3.0, 5.0, 7.0, 8.5, 9.0, and 9.5; $\times 10^5 \text{ s}^{-1}$), and moist static energy (blue contours, at intervals of 350, 351, 352, and 353; $\times 10^{-3} \text{ J kg}^{-1}$) in (a) NOAH_WSM6, (b) RUC_WSM6, (c) NOAH_MOR, and (d) RUC_MOR. The numbers on the y axis denote the approximate height AGL (m) at different η vertical levels.

where

$$\text{PV} = -\frac{g}{\mu} \left[(\zeta_\eta + f) \frac{\partial \theta}{\partial \eta} + \left(\frac{\partial u}{\partial \eta} \frac{\partial \theta}{\partial y} - \frac{\partial v}{\partial \eta} \frac{\partial \theta}{\partial x} \right) \right], \quad (2)$$

$$\widetilde{\omega}_\eta = -\frac{1}{\mu} \int_0^\eta \nabla(\mu \mathbf{V}_h) d\eta + \frac{\eta}{\mu} \int_0^1 \nabla(\mu \mathbf{V}_h) d\eta, \quad \text{and} \quad (3)$$

$$\mu = p_s - p_t. \quad (4)$$

In the above equations, $\mathbf{V}_h$ denotes the horizontal wind vectors (m s^{-1}); u , v , $\widetilde{\omega}_\eta$, θ , H , ζ_η , g , p_s , and p_t represent zonal wind (m s^{-1}), meridional wind (m s^{-1}), vertical velocity in the η vertical coordinate (s^{-1}), potential temperature (K), diabatic heating (K s^{-1}), vertical relative vorticity in the η vertical coordinate (s^{-1}), gravitational acceleration (m s^{-2}), surface pressure (Pa), and model-top pressure (equal to 500 Pa in the default WRF setting), respectively.

Although Eq. (1) is useful for estimating the spatial structures of PV budgets in TPV, it would be more appropriate to use the area- or volume-integrated PV

budgets that can distinguish internal PV forcing processes from boundary PV fluxes. Similar to Kieu and Zhang (2009), Eq. (1) is written in area- or volume-integration form and then averaged by the specific area or volume, hereinafter referred to as area-integrated PV budgets (APV) or volume (bulk)-integrated PV budgets (BPV); that is,

$$\int_V \frac{\partial \text{PV}}{\partial t} dV = - \int_V \nabla_\eta (\mathbf{V}_3 \cdot \text{PV}) dV + \int_V (\text{PV} \cdot \nabla_\eta \mathbf{V}_3) dV - \int_V \left(\frac{g}{\mu} \xi_{a\eta} \cdot \nabla_\eta H \right) dV, \quad (5)$$

where $\mathbf{V}_3$ denotes the three-dimensional wind vectors in the η vertical coordinate (m s^{-1}), $\xi_{a\eta} = \zeta_\eta + f$ is the vertical absolute vorticity in the η vertical coordinate (s^{-1}), and V is a specific area or volume. Specifically, the area is defined as $30^\circ\text{--}34^\circ\text{N}$, $85^\circ\text{--}93^\circ\text{E}$; the volume in its horizontal plane is the area of $30^\circ\text{--}34^\circ\text{N}$, $85^\circ\text{--}93^\circ\text{E}$ and in its vertical levels, defined as $8 \leq \eta \leq 24$ (about 1100–8826 m AGL) to neglect the influence of

FIG. 10. As in Fig. 9, but for microphysical particles regarding the mixing ratio (g kg^{-1}) of ice (QICE; color shaded), cloud (QCLD; gray shaded), vapor (QVAPOR; blue contours, at intervals of 6.5 and 7.5), snow (QSNOW; black contours, at intervals of 0.04, 0.08, and 0.12) and graupel (QGRAUP; red contours, at an interval of 0.04).

friction and to minimize the impact of high PV in the stratosphere since the high PV in the lower atmosphere and stratosphere are clearly separated by the low PV in the middle atmosphere; that is, the local increase in PV associated with nighttime TPV cannot be due to an intrusion of stratospheric air via subsidence (not shown). The ∇_η and $\xi_{a\eta}$ denote the divergence operator and absolute vorticity in the η vertical coordinate, and the other notations have been explained above. The diabatic heating term H in Eq. (5) is condensational latent heating derived directly from cloud microphysics schemes.

Equation (5) states that the time rates of APV (BPV) changes $[\int_V (\partial \text{PV} / \partial t) dV; \text{QTND}]$ are determined by the terms on its right-hand side, which are from left to right the net cross-boundary PV fluxes $[-\int_V \nabla_\eta (\mathbf{V}_3 \cdot \text{PV}) dV; \text{QBND}]$, the condensing or diluting rates of APV (BPV) due to the 3D velocity divergence $[\int_V (\text{PV} \cdot \nabla_\eta \mathbf{V}_3) dV; \text{QCON}]$, and the diabatic PV production rates $\{-\int_V [(g/\mu) \xi_{a\eta} \cdot \nabla_\eta H] dV; \text{QDBH}\}$. The BPV and APV represent area and volume-integrated PV budgets, which are used in the analysis in Figs. 11 and 12 below, respectively.

Figure 11 compares the time series of the BPV budget in the different experiments. Results indicate that the forcing term QDBH3d associated directly with diabatic heating shows time rates of changes that are opposite to QBND3d, namely, with strong positive contributions during the daytime but decreased or even negative contributions during the nighttime. The contributions of QCON3d to QTND are not evident in any of the simulations. In addition, the stronger contributions of daytime QDBH3d and nighttime QBND3d are conducive to the increase in BPV and thus have positive impacts on TPV genesis. The intermittency of vortices at different scales entering the control volume, as well as the large variability in diabatic heating gradients associated with convections over the TP cause QTND, QBND3d, QCON3d, and QDBH3d exhibit marked fluctuations.

To obtain further in-depth physical insight into the processes associated with the genesis of nighttime TPV, diagnosis of daytime and nighttime averaged vertical profiles of the APV budget is analyzed in Fig. 12 to quantify the relative contributions of each forcing term to the APV tendency. In particular, the

FIG. 11. Time series of the volume-integrated PV budget ($\text{BPV}; \times 10^6 \text{ PVU s}^{-1}$) in the different experiments.

horizontal components of each APV budget on the right-hand side of Eq. (5) are also presented. Results show that during both the daytime and nighttime, QTND is positive in the lower atmosphere (below 3500 AGL) but is almost zero in the middle and upper atmosphere, indicating that the local increase in APV associated with the nighttime TPV cannot be related to the intrusion of stratospheric air via subsidence. Since positive vorticity associated with the TPV is located mainly in the lower atmosphere, therefore, we will focus on the APV budget analysis mainly in the lower atmosphere here.

During the daytime, QDBH3d has strong positive contributions to QTND because the vertical component of QDBH is positive and tends to dominate its negative horizontal component; also during the daytime, QBND3d has strong negative contributions to QTND because the vertical component of QBND is negative and tends to dominate its positive horizontal component. Although the positive vertical component of QCON and the negative horizontal component of QCON offset each other, the latter is relatively weaker, resulting in the weaker positive contribution of QCON3d to QTND. Stronger condensational latent heating corresponds well with the stronger positive contribution of QDBH3d to QTND, and combined with the results in Figs. 8 and 9, this indicates that stronger daytime surface diabatic heating is the main mechanism for PV increase. The intensity of daytime surface diabatic heating is essential to the development of daytime convections. During the nighttime, QTND increases obviously in the lower atmosphere,

especially in RUC_MOR, indicating that the cyclonic vorticity associated with TPV genesis strengthens and is strongest in RUC_MOR, consistent with the results in Figs. 5 and 9. Meanwhile, it is found that the strengthened QTND in the lower atmosphere could be because of two reasons: 1) The negative or weak positive QBND3d during the daytime increases obviously and becomes positive during the nighttime, although QBND2d is very similar during both the daytime and nighttime, indicating that the strengthened and positive vertical component of QBND during the nighttime has important and positive impacts on the PV increase in the lower atmosphere; in other words, the TPV forms as a result of merging convections, with the vertical component of the net cross-boundary PV fluxes playing an important role. 2) Condensational latent heating decreases are not obvious during the nighttime (also can be seen in Fig. 9), especially when daytime surface diabatic heating is stronger, indicating that strong daytime surface diabatic heating is conducive to convection development and thus could guarantee the existence of strong nighttime condensational latent heating. Meanwhile, the Morrison scheme tends to produce stronger condensational latent heating. As a result, condensational latent heating in the lower atmosphere is strongest in RUC_MOR, leading to the most significant positive contribution of QDBH3d to the PV increase (e.g., about 1100–1500 m AGL). Therefore, the strongest cyclonic vorticity associated with TPV genesis is simulated in RUC_MOR.

FIG. 12. Vertical profiles of the area-integrated PV budget (APV; $\times 10^6$ PVU s^{-1}) and condensational latent heating (1.5×10^5 K s^{-1}) during (top) the daytime from 0600 UTC 28 Jul to 1200 UTC 28 Jul and (bottom) the nighttime from 1400 UTC 28 Jul to 0000 UTC 29 Jul in (a),(e) NOAH_WSM6, (b),(f) RUC_WSM6, (c),(g) NOAH_MOR, and (d),(h) RUC_MOR.

Overall, the merging of convections and the condensational latent heating both have important impacts on the generation of the closed cyclonic vorticity associated with the nighttime TPV. The strong daytime surface diabatic heating could provide a favorable condition for the nighttime TPV genesis.

5. Concluding remarks

This study investigates the impacts of surface diabatic and atmospheric condensational latent heating on TPV genesis in a diurnal cycle, using a series of numerical sensitivity experiments. It is found that the genesis of nighttime TPV is closely related to the intensity of daytime convections since it forms as a result of merging convections. Compared to observations, the WRF Model can well reproduce the evolution of daytime convections and nighttime TPV with respect to cloud-top brightness temperature, surface rainfall patterns, etc., under a specific parameterization configuration. However, colder and drier biases in the lower atmosphere, and

drier biases in the upper atmosphere, which degrade the simulation performance, are still present, especially during the daytime.

Intercomparisons among the experiments by combining Noah and RUC land surface schemes and WSM6 and Morrison cloud microphysics schemes in terms of mid- and low-level cyclonic vorticity, surface diabatic and condensational latent heating, moist static energy, and microphysical particles indicate that the development of convections is more favorable when daytime surface diabatic heating is vigorous. Moreover, the simulation discrepancies in these quantities are more significant with different land surface schemes than with different cloud microphysics schemes. The simulation results from different cloud microphysics schemes and their associated discrepancies are largely influenced by the choice of land surface scheme, highlighting that surface diabatic heating induced by the land surface scheme plays a dominant role in the development of daytime convection and the genesis of the nighttime TPV.

Further diagnosis of the PV budget indicates that the obvious increase in PV in the lower atmosphere during the nighttime, especially in the experiment consists of RUC land surface scheme and Morrison cloud microphysics scheme, could be portrayed by evidently strengthened cyclonic vorticity associated with TPV genesis. The increased PV could be attributed to the increased vertical component of net cross-boundary PV fluxes during the merging of convections, as well as the significant positive contribution of diabatic heating effects due to the strong nighttime condensational latent heating in the lower atmosphere. Since the genesis of the nighttime TPV is closely related to the intensity of daytime convections, it is concluded that strong daytime surface diabatic heating, which is essential to the development of daytime convections, could provide a favorable condition for nighttime TPV genesis.

The framework proposed in this study is expected to be useful for understanding the role of surface diabatic and condensational latent heating in TPV genesis and the thermodynamic mechanisms related to TPV genesis. However, it should be noted that the results of this study should not be used as a commentary on which parameterization or experiment design is better, since the conclusions derived from this study are based on only one TPV case that is generated and dissipates in its origins over the TP, and our motivation is to understand the thermodynamic mechanisms of TPV genesis. Moreover, previous studies (e.g., Cui and Wang 2009; Bao and Zhang 2013) found that different reanalysis datasets generally have biases and great uncertainty over the TP, especially in the lower atmosphere. Future work should focus on the improvement of initial conditions and physics parameterizations such as land surface, boundary layer, and cloud microphysics processes, especially in the lower atmosphere and over the complex terrain (Pu et al. 2013; Zhang and Pu 2017; Zhang et al. 2015, 2017, 2019), based on intense observations of more TPV cases. The uncertainties of complex terrain over the TP on the simulation of TPV should also be the interesting topic that deserves to be further investigated.

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REFERENCES

- Bao, X., and F. Zhang, 2013: Evaluation of NCEP–CFSR, NCEP–NCAR, ERA–Interim, and ERA–40 Reanalysis datasets against independent sounding observations over the Tibetan Plateau. *J. Climate*, **26**, 206–214, <https://doi.org/10.1175/JCLI-D-12-00056.1>.
- Benjamin, S. G., G. A. Grell, J. M. Brown, T. G. Smirnova, and R. Bleck, 2004: Mesoscale weather prediction with the RUC hybrid isentropic–terrain-following coordinate model. *Mon. Wea. Rev.*, **132**, 473–494, [https://doi.org/10.1175/1520-0493\(2004\)132<0473:MWPWTR>2.0.CO;2](https://doi.org/10.1175/1520-0493(2004)132<0473:MWPWTR>2.0.CO;2).
- Chen, F., and J. Dudhia, 2001: Coupling an advanced land surface hydrology model with the Penn State–NCAR MM5 modeling system. Part I: Model implementation and sensitivity. *Mon. Wea. Rev.*, **129**, 569–585, [https://doi.org/10.1175/1520-0493\(2001\)129<0569:CAALSH>2.0.CO;2](https://doi.org/10.1175/1520-0493(2001)129<0569:CAALSH>2.0.CO;2).
- Cui, Y., and C. Wang, 2009: Comparison of sensible and latent heat fluxes during the transition season over the western Tibetan Plateau from reanalysis datasets. *Prog. Nat. Sci.*, **19**, 719–726, <https://doi.org/10.1016/j.pnsc.2008.11.001>.
- Curio, J., R. Schiemann, K. Hodges, and A. Turner, 2019: Climatology of Tibetan Plateau vortices in reanalysis data and a high-resolution global climate model. *J. Climate*, **32**, 1933–1950, <https://doi.org/10.1175/JCLI-D-18-0021.1>.
- Dell’Osso, L., and S. J. Chen, 1986: Numerical experiments on the genesis of vortices over the Qinghai–Tibet plateau. *Tellus*, **38A**, 236–250, <https://doi.org/10.3402/tellusa.v38i3.11715>.
- Eltahir, E. A., 1998: A soil moisture–rainfall feedback mechanism: 1. Theory and observations. *Water Resour. Res.*, **34**, 765–776, <https://doi.org/10.1029/97WR03499>.
- Feng, X., C. Liu, R. Rasmussen, and G. Fan, 2014: A 10-yr climatology of Tibetan Plateau vortices with NCEP Climate Forecast System reanalysis. *J. Appl. Meteor. Climatol.*, **53**, 34–46, <https://doi.org/10.1175/JAMC-D-13-014.1>.
- Gao, Y., J. Xu, and D. Chen, 2015: Evaluation of WRF mesoscale climate simulations over the Tibetan Plateau during 1979–2011. *J. Climate*, **28**, 2823–2841, <https://doi.org/10.1175/JCLI-D-14-00300.1>.
- Giorgi, F., L. O. Mearns, C. Shields, and L. Mayer, 1996: A regional model study of the importance of local versus remote controls of the 1988 drought and the 1993 flood over the central United States. *J. Climate*, **9**, 1150–1162, [https://doi.org/10.1175/1520-0442\(1996\)009<1150:ARMSOT>2.0.CO;2](https://doi.org/10.1175/1520-0442(1996)009<1150:ARMSOT>2.0.CO;2).
- Hong, S. Y., J. Dudhia, and S. H. Chen, 2004: A revised approach to ice microphysical processes for the bulk parameterization of clouds and precipitation. *Mon. Wea. Rev.*, **132**, 103–120, [https://doi.org/10.1175/1520-0493\(2004\)132<0103:ARATIM>2.0.CO;2](https://doi.org/10.1175/1520-0493(2004)132<0103:ARATIM>2.0.CO;2).
- , Y. Noh, and J. Dudhia, 2006: A new vertical diffusion package with an explicit treatment of entrainment processes. *Mon. Wea. Rev.*, **134**, 2318–2341, <https://doi.org/10.1175/MWR3199.1>.
- Iacono, M. J., J. S. Delamere, E. J. Mlawer, M. W. Shephard, S. A. Clough, and W. D. Collins, 2008: Radiative forcing by long-lived greenhouse gases: Calculations with the AER radiative transfer models. *J. Geophys. Res.*, **113**, D13103, <https://doi.org/10.1029/2008JD009944>.
- Jiménez, P. A., J. Dudhia, J. F. González-Rouco, J. Navarro, J. P. Montávez, and E. García-Bustamante, 2012: A revised scheme

- for the WRF surface layer formulation. *Mon. Wea. Rev.*, **140**, 898–918, <https://doi.org/10.1175/MWR-D-11-00056.1>.
- Kain, J. S., 2004: The Kain–Fritsch convective parameterization: An update. *J. Appl. Meteor.*, **43**, 170–181, [https://doi.org/10.1175/1520-0450\(2004\)043<0170:TKCPAU>2.0.CO;2](https://doi.org/10.1175/1520-0450(2004)043<0170:TKCPAU>2.0.CO;2).
- Kieu, C. Q., and D. L. Zhang, 2009: Genesis of Tropical Storm Eugene (2005) associated with the ITCZ breakdowns. Part II: Roles of vortex merger and ambient potential vorticity. *J. Atmos. Sci.*, **66**, 1980–1996, <https://doi.org/10.1175/2008JAS2905.1>.
- Koshyk, J. N., and N. A. McFarlane, 1996: The potential vorticity budget of an atmospheric general circulation model. *J. Atmos. Sci.*, **53**, 550–563, [https://doi.org/10.1175/1520-0469\(1996\)053<0550:TPVBOA>2.0.CO;2](https://doi.org/10.1175/1520-0469(1996)053<0550:TPVBOA>2.0.CO;2).
- Lhasa Group for Tibetan Plateau Meteorology Research, 1981: *Research of 500 hPa Shear Lines over the Tibetan Plateau in Summer* (in Chinese). Science Press, 122 pp.
- Li, G. P., H. G. Lu, C. H. Huang, Y. Y. Fan, and B. Zhang, 2016: A climatology of the surface heat source on the Tibetan Plateau in summer and its impacts on the formation of the Tibetan Plateau vortex (in Chinese). *Chin. J. Atmos. Sci.*, **40**, 131–141.
- Li, L., R. Zhang, M. Wen, and L. Liu, 2014: Effect of the atmospheric heat source on the development and eastward movement of the Tibetan Plateau vortices. *Tellus*, **66A**, 24451, <https://doi.org/10.3402/tellusa.v66.24451>.
- Liu, Y. F., and G. P. Li, 2016: Climatic characteristics of atmospheric heat source over the Tibetan Plateau and its possible relationship with the generation of the Tibetan Plateau vortex in the summer (in Chinese). *Chin. J. Atmos. Sci.*, **40**, 864–876.
- Luo, S. W., M. L. He, and X. D. Liu, 1993: Study on summer vortices over Qinghai–Xizang Plateau (in Chinese). *Sci. China*, **23B**, 778–784.
- Maussion, F., D. Scherer, R. Finkelnburg, J. Richters, W. Yang, and T. Yao, 2011: WRF simulation of a precipitation event over the Tibetan Plateau, China—An assessment using remote sensing and ground observations. *Hydrol. Earth Syst. Sci.*, **15**, 1795–1817, <https://doi.org/10.5194/hess-15-1795-2011>.
- , —, T. Mölg, E. Collier, J. Curio, and R. Finkelnburg, 2014: Precipitation seasonality and variability over the Tibetan Plateau as resolved by the High Asia Reanalysis. *J. Climate*, **27**, 1910–1927, <https://doi.org/10.1175/JCLI-D-13-00282.1>.
- Morrison, H., G. Thompson, and V. Tatarskii, 2009: Impact of cloud microphysics on the development of trailing stratiform precipitation in a simulated squall line: Comparison of one- and two-moment schemes. *Mon. Wea. Rev.*, **137**, 991–1007, <https://doi.org/10.1175/2008MWR2556.1>.
- , J. A. Milbrandt, G. H. Bryan, K. Ikeda, S. A. Tessendorf, and G. Thompson, 2015: Parameterization of cloud microphysics based on the prediction of bulk ice particle properties. Part II: Case study comparisons with observations and other schemes. *J. Atmos. Sci.*, **72**, 312–339, <https://doi.org/10.1175/JAS-D-14-0066.1>.
- Pu, Z., H. L. Zhang, and J. A. Anderson, 2013: Ensemble Kalman filter assimilation of near-surface observations over complex terrain: Comparison with 3DVAR for short-range forecasts. *Tellus*, **65A**, 19620, <https://doi.org/10.3402/tellusa.v65i0.19620>.
- Qiao, Q. M., 1987: The environment analysis on 500hPa vortices moving eastward out of Tibet Plateau in summer (in Chinese). *Plateau Meteor.*, **6**, 45–54.
- Shen, R., E. R. Reiter, and J. F. Bresch, 1986a: Some aspects of the effects of sensible heating on the development of summer weather systems over the Tibetan Plateau. *J. Atmos. Sci.*, **43**, 2241–2260, [https://doi.org/10.1175/1520-0469\(1986\)043<2241:SAOTEO>2.0.CO;2](https://doi.org/10.1175/1520-0469(1986)043<2241:SAOTEO>2.0.CO;2).
- , —, and —, 1986b: Numerical simulation of the development of vortices over the Qinghai–Xizang (Tibet) Plateau. *Meteor. Atmos. Phys.*, **35**, 70–95, <https://doi.org/10.1007/BF01029526>.
- Shou, Y., F. Lu, H. Liu, P. Cui, S. Shou, and J. Liu, 2019: Satellite-based observational study of the Tibetan Plateau vortex: Features of deep convective cloud tops. *Adv. Atmos. Sci.*, **36**, 189–205, <https://doi.org/10.1007/s00376-018-8049-y>.
- Skamarock, W. C., and Coauthors, 2008: A description of the Advanced Research WRF version 3. NCAR Tech. Note NCAR/TN-475/STR, 113 pp., <https://doi.org/10.5065/D68S4MVH>.
- Sugimoto, S., and K. Ueno, 2010: Formation of mesoscale convective systems over the eastern Tibetan Plateau affected by plateau-scale heating contrasts. *J. Geophys. Res.*, **115**, D16105, <https://doi.org/10.1029/2009JD013609>.
- Tian, S., A. Duan, Z. Wang, and Y. Gong, 2015: Interaction of surface heating, the Tibetan Plateau vortex, and a convective system: A case study (in Chinese). *Chin. J. Atmos. Sci.*, **39**, 125–136.
- Wan, B., Z. Gao, F. Chen, and C. Lu, 2017: Impact of Tibetan Plateau surface heating on persistent extreme precipitation events in southeastern China. *Mon. Wea. Rev.*, **145**, 3485–3505, <https://doi.org/10.1175/MWR-D-17-0061.1>.
- Wang, B., 1987: The development mechanism for Tibetan Plateau warm vortices. *J. Atmos. Sci.*, **44**, 2978–2994, [https://doi.org/10.1175/1520-0469\(1987\)044<2978:TDMFTP>2.0.CO;2](https://doi.org/10.1175/1520-0469(1987)044<2978:TDMFTP>2.0.CO;2).
- , and I. Orlanski, 1987: Study of a heavy rain vortex formed over the eastern flank of the Tibetan Plateau. *Mon. Wea. Rev.*, **115**, 1370–1393, [https://doi.org/10.1175/1520-0493\(1987\)115<1370:SOAHRV>2.0.CO;2](https://doi.org/10.1175/1520-0493(1987)115<1370:SOAHRV>2.0.CO;2).
- Wang, C., D. Wu, and F. Zhang, 2019: Modification of the convective adjustment time-scale in the Kain–Fritsch eta scheme for the case of weakly forced deep convection over the Tibetan Plateau region. *Quart. J. Roy. Meteor. Soc.*, **145**, 1915–1932, <https://doi.org/10.1002/qj.3535>.
- Wu, D., F. Zhang, and C. Wang, 2018: Impacts of diabatic heating on the genesis and development of an inner Tibetan Plateau vortex. *J. Geophys. Res.*, **123**, 11 691–11 704, <https://doi.org/10.1029/2018JD029240>.
- Xiang, S., Y. Li, D. Li, and S. Yang, 2013: An analysis of heavy precipitation caused by a retracing plateau vortex based on TRMM data. *Meteor. Atmos. Phys.*, **122**, 33–45, <https://doi.org/10.1007/s00703-013-0269-1>.
- Xu, W., and Y. Zhang, 2017: Numerical study on the feedback between latent heating and convection in a Qinghai–Tibetan Plateau vortex. *Plateau Meteor.*, **36**, 763–775.
- Ye, D. Z., and Y. X. Gao, 1979: *The Tibetan Plateau Meteorology* (in Chinese). Science Press, 278 pp.
- Yu, S., W. Gao, D. Xiao, and J. Peng, 2016: Observational facts regarding the joint activities of the southwest vortex and plateau vortex after its departure from the Tibetan Plateau. *Adv. Atmos. Sci.*, **33**, 34–46, <https://doi.org/10.1007/s00376-015-5039-1>.
- Zeng, X., and Coauthors, 2016: Effects of land surface schemes on WRF-simulated geopotential heights over China in summer 2003. *J. Hydrometeorol.*, **17**, 829–851, <https://doi.org/10.1175/JHM-D-14-0239.1>.
- Zhang, F., and Z. Pu, 2017: Effects of vertical eddy diffusivity parameterization on the evolution of landfalling hurricanes. *J. Atmos. Sci.*, **74**, 1879–1905, <https://doi.org/10.1175/JAS-D-16-0214.1>.

- , Y. Yang, and C. Wang, 2015: The effects of assimilating conventional and ATOVS data on forecasted near-surface wind with WRF-3DVAR. *Mon. Wea. Rev.*, **143**, 153–164, <https://doi.org/10.1175/MWR-D-14-00038.1>.
- , Z. Pu, and C. Wang, 2017: Effects of boundary layer vertical mixing on the evolution of hurricanes over land. *Mon. Wea. Rev.*, **145**, 2343–2361, <https://doi.org/10.1175/MWR-D-16-0421.1>.
- , —, and —, 2019: Impacts of soil moisture on the numerical simulation of a post-landfall storm. *J. Meteor. Res.*, **33**, 206–218, <https://doi.org/10.1007/s13351-019-8002-8>.
- Zhang, P., G. P. Li, X. Fu, Y. Liu, and L. Li, 2014: Clustering of Tibetan Plateau vortices by 10–30-day intraseasonal oscillation. *Mon. Wea. Rev.*, **142**, 290–300, <https://doi.org/10.1175/MWR-D-13-00137.1>.
- Zulkaflı, Z., W. Buytaert, C. Onof, B. Manz, E. Tarnavsky, W. Lavado, and J. Guyot, 2014: A comparative performance analysis of TRMM 3B42 (TMPA) versions 6 and 7 for hydrological applications over Andean–Amazon River basins. *J. Hydrometeor.*, **15**, 581–592, <https://doi.org/10.1175/JHM-D-13-094.1>.