Characteristics of Summer Convective Systems Initiated over the Tibetan Plateau. Part I: Origin, Track, Development, and Precipitation

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ABSTRACT

Summer convective systems (CSs) initiated over the Tibetan Plateau identified by the International Satellite Cloud Climatology Project (ISCCP) deep convection database and associated Tropical Rainfall Measuring Mission (TRMM) precipitation for 1998–2001 have been analyzed for their basic characteristics in terms of initiation, distribution, trajectory, development, life cycle, convective intensity, and precipitation. Summer convective systems have a dominant center over the Hengduan Mountain and a secondary center over the Yangtze River Valley. Precipitation associated with these CSs contributes more than 60% of total precipitation over the central-eastern area of the Tibetan Plateau and 30%–40% over the adjacent region to its southeast. The average CS life cycle is about 36 h; 85% of CSs disappear within 60 h of their initiation. About 50% of CSs do not move out of the Tibetan region, with the remainder split into eastward- and southward-moving components. These CSs moving out of the Tibetan Plateau are generally larger, have longer life spans, and produce more rainfall than those staying inside the region. Convective system occurrences and associated rainfall present robust diurnal variations. The midafternoon maximum of CS initiation and associated rainfall over the plateau is mainly induced by solar heating linked to the unique Tibetan geography. The delayed afternoon–late night peak of rainfall from CSs propagating out of this region is a combined outcome of multiple mechanisms working together. Results suggest that interactions of summer Tibetan CSs with the orientation of the unique Tibetan geography and the surrounding atmospheric circulations are important for the development, intensification, propagation, and life span of these CSs.

1. Introduction

The effect of the Tibetan Plateau on the atmospheric systems has been studied in detail for several decades (Reiter 1982), in terms of boundary layer, physical processes of the surface–air interactions, heating sources, meteorological elements, synoptic weather systems, general atmospheric circulations, and climate variations. Various methods have been utilized in the processes, such as boundary layer observational analysis (Yanai and Li 1994; Gao et al. 2002), objective analyses of synoptic weather and general atmospheric circulation (Ye 1981; Yanai et al. 1992), statistical study of synoptic weather systems (Tao and Ding 1981; Leber et al. 1995), analysis of satellite measurements (Jin 1997; Ueno 1998), diagnostic studies of meteorological vari-
ables and atmospheric energy (Dell’Osso and Chen 1986; Reiter 1987), numerical simulations (Shen et al. 1986b), fluid dynamic experiments (Boyer and Chen 1987), and theoretic studies of dynamics (Xie 1981; Le- roux 1993). The ultimate goal of these investigations is to describe the processes of thermodynamics and dynamics of the Tibetan Plateau, and to understand their impact on global climate change. Results could lead to a new conceptual model and a prediction theory for severe weather systems influenced by the plateau, with further understanding of the environmental conditions of the Tibetan synoptic systems and climate, and the characteristics of structure and movement of Tibetan synoptic systems as well as the Tibetan heating and topographical forcing.

It is well known that the Tibetan Plateau is a source of dynamic and thermodynamic turbulence (Zhao and Chen 2000). Many severe weather systems that impacted China in the past are linked to the dynamic and thermodynamic influences of the plateau (Tao and Ding 1981; Reiter 1982; Shen et al. 1986a,b). Tao et al. (1980) indicate that heavy rainfall in China is mainly from typhoon, midlatitude frontal systems, and cyclonic vortices that propagate eastward from Tibet. Jiang et al. (1996) demonstrate that the eastward propagation of convective cloud systems over the Tibet Plateau could trigger heavy precipitation over the Yangtze River basin. For example, the persistent extreme heavy rainfall and flooding over the Yangtze and Huihe River basins in 1991 and the Yangtze River basin in 1998 were primarily caused by the cyclonic vortices that originated from Tibet (Fang 1985; Li et al. 1989; Xiang and Jiang 1995; Shi et al. 2000; Jiang and Fan 2002). The surface heating over the Tibetan Plateau is an elevated heating source in the midtroposphere (Luo and Yanai 1983, 1984; Yanai et al. 1992; Yanai and Li 1994). Wu et al. (1997, 2002a,b), Wu and Liu (2000, 2003), and Liu et al. (2001, 2002) describe in detail the characteristics of the summer atmosphere general circulation forced by the Tibetan heating, and propose a mechanism to explain how the atmospheric circulation is influenced by the Tibetan heating based on the theory of thermodynamic geostrophic adjustment. This mechanism is defined as the “sensible heat driven air pump.” They demonstrate that positive vorticity would be generated below the altitude of the Tibetan heating source, where heating intensity increases with height. A negative vorticity would be produced above the heating source, where the heating rate decreases with height. The low-level positive vorticity will balance the upper-level negative vorticity and the transport of negative vorticity from the Tibetan lateral boundaries, and depletion from surface friction. A deep anticyclone vortex would be induced by the horizontally inhomogeneous heating distribution in the north of the heating source with westerly existence. Wu and Zhang (1998) also suggest that the onset of the Asian summer monsoon mainly depends on the Tibet heating source. Furthermore, results based on numerical model simulations by He et al. (1984), Wang et al. (1984), and Qian et al. (1988) show that there would be no Tibetan high pressure center, no breakup of the west Pacific subtropical high, and no tube of monsoon circulation without the Tibetan heating source. In addition, there would be no convergence zone in eastern Tibet during the time period of mei-ju. Therefore, the physical processes and atmospheric circulation of the plateau have important impacts on global climate, Asian atmospheric circulation, abnormal weather phenomena, and the climatology of severe weather.

Literature results have shown that strong convective weather systems (such as heavy rainfall, ice storms, tornadoes, wind gusts, and downdraft flows) are mostly associated with mesoscale convective systems (MCSs) (McCollum et al. 1995; Gray 2001). Because of its unique thermodynamic forcing, the Tibetan Plateau is an active region of strong summer CSs (Shen et al. 1986a,b; Wang et al. 1993). Flohn (1968) demonstrates 20–50 well-developed cumulus clouds per 10, 000 km² over the region, which indicates frequent summer convective activity. Fu et al. (2006) show that these strong convective systems play a central role in transporting water vapor to the global stratosphere during summer seasons. Thus, the characterization of these CSs is also important in understanding the global stratosphere composition and climate.

Jiang and Fan (2002) indicate two maximum convection centers separated around 95°E over the region (i.e., the southeast center and the relatively stronger south center). Only a small portion of these CSs propagates eastward, influencing precipitation over the Yangtze River basin. Although only a few eastward CSs intensify as they propagate, they often trigger heavy rainfall events over the middle low segments of the Yangtze River basin. (The eastward propagation of Tibetan CSs or MCSs is for a system moving east of the Tibetan Plateau during its life cycle.) Zhuo et al. (2002) point out that these CSs tend to propagate eastward toward the mid-low segments of the Yangtze River basin, where the occurrence of heavy rainfall events were generally coincident with the arrivals of these CSs. The eastward propagation of a CS appears to be in favor of following large-scale dynamic circulation patterns: 1) the fields of isobar height, divergence, and vorticity at 400 hPa show a south–north orientation pattern; and 2) the atmospheric thermo-instability index and isobar
height at 500 hPa has a west–east orientation distribution (Guo et al. 2003a,b, 2004; Fang et al. 2004a,b).

However, most previous investigations of the Tibetan CS or MCS were case studies, focusing on the vicinity of the Plateau. The characteristics of the general atmospheric circulations during CS propagation and development, and the relationship between propagation and environmental variables, remain unclear. Therefore, the basic characteristics of these CSs and their propagation, development, life cycles, and impacts on surface rainfall are the main objectives of this study. We focus on the mean properties of CSs initiated over the Tibetan Plateau in terms of intensity, size, propagation, and precipitation, by tracking all summer CSs and associated surface rainfall from satellite measurements in 1998–2001.

2. Methodology and datasets

a. Datasets

There are only a few standard meteorological stations over the Tibetan Plateau because of its harsh climate and unique geography features, and high-quality traditional meteorological datasets with a long time series and large area coverage are lacking for this region. In addition, ground radar remote sensing cannot provide good measurements over the remote mountain area. However, satellite remote sensing can overcome those issues associated with the traditional observation techniques. The International Satellite Cloud Climatology Project (ISCCP) Convection Tracking Database (CTD) is utilized to investigate the summer CSs originating from the Tibetan region. Since Tropical Rainfall Measuring Mission (TRMM) precipitation products are to date one of the best rainfall datasets (Kummerow et al. 2001; Yang et al. 2006a), the TRMM blended rainfall (3B42) and the precipitation radar (PR)–based rainfall (2A25) rain products are used in this study (Iguchi et al. 2000; Meneghini et al. 2001; Huffman et al. 2001, 2007).

CTD has 41 parameters related to the macrocharacteristics and cloud physics of CSs during their life cycles. It systematically provides descriptions of CSs in terms of propagation, track, speed, intensity, location of heavy rainfall, internal cloud properties, and size. CTD is constructed using the following procedures (Rossow et al. 1996): first, clear or cloudy pixels are identified using the infrared (IR) and visible (VIS) observations. The VIS data are only applied during daytime. If cloudy, the cloud-top brightness temperature (TB) and cloud optical depth (daytime only) are obtained. This information is used to build up the ISCCP DX database (Rossow et al. 1996; Rossow and Schiffer 1999). The DX dataset has 3-h temporal resolution and associated pixels at 30 km × 30 km spatial resolution. Second, analyzing the DX database leads to CTD. This procedure includes two steps (i.e., identification of cloud clusters and tracking analysis). The identification of cloud clusters from satellite images follows the method proposed by Machado and Rossow (1993); that is, a CS is defined as adjacent cloud clusters where TB < 245 K, while a convective cluster (CC) is determined as adjacent cloud clusters where TB < 220 K. Thus, a CC is a relatively stronger convective cell than a CS. These procedures are consistent with the published literature (Clark 1983; Maddox 1980; Machado et al. 1998). In general, there are several CCs inside the coverage of a CS. The tracking analysis based on the method by Machado et al. (1998) has two substeps: 1) determination of the time interval between two continuous images; and 2) determination of the overlap area of a convective system in two continuous images by calculating the ratio between all pixels in the overlapped region of two convective systems and the total pixels of these two convective systems. This process is conducted for all possible pairs of overlapped convective systems. The pair with the largest ratio is selected as the convective system to be tracked. Two more tests are performed to assess the quality of the tracking analysis. In case of missing satellite images, the CS tracking continues if the time separation is less than the time criterion estimated with the CS’s size (Machado et al. 1998). The CS for this time step is then considered as missing so that it would not be counted in the statistical analysis. Because the situation of missing images happens only occasionally, it would not significantly impact the results of this study. The CTD includes all convective systems with a minimum radius of 90 km (corresponding to a minimum of 30 pixels in the DX dataset) over the global 70°S–70°N belt. The datasets are available at 0000, 0300, 0600, 0900, 1200, . . . , 2100 UTC. Detailed documentation on the ISCCP dataset can be found online (http://isccp.giss.nasa.gov/).

The initial observations contributing to the CTD dataset come from five geostationary satellites: the U.S. Geostationary Operational Environmental Satellite (GOES) East (GOE), GOES-West (GOW), Japanese Geostationary Meteorological Satellite (GMS), Europe Meteorological Satellite (MET), and India Satellite (INS).

The TRMM 3B42 rain product based on rain retrievals from multiple satellite passive microwave and IR measurements is available at 0000, 0300, 0600, 0900, . . . , 2100 UTC at 0.25° × 0.25° grid resolution
This product ultimately utilizes high-accuracy rain estimates from satellite passive microwave observations [from TRMM, Special Sensor Microwave Imager (SSM/I), Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E), and Advanced Microwave Sounding Unit (AMSU-B)] and high temporal resolution from satellite IR measurements. Any valid rain rates from available passive microwave measurements within 90 min of a designated data point are selected as the rain rate for this point. If no passive microwave–based rain rate is available, the rain rate from IR measurements adjusted with passive microwave–based rain rates is used for this data point. The 3B42 monthly rain estimates are finally adjusted by surface rain gauge measurements. Published results demonstrate that the 3B42 rain products are highly comparable to any other satellite remotely sensed, surface gauge, and radar rain products (Huffman et al. 2007; E. A. Smith et al. 2008, unpublished manuscript). However, it was noted that 3B42 rain locations could shift a little bit at storm edge because of the IR rain estimates. In addition, E. A. Smith et al. (2008, unpublished manuscript) indicate that the surface gauge adjustments might lead to the underestimation of 3B42 rainfall over the west U.S. mountain regions during the warm season because of less reliable surface rain gauges over these mountain areas and the evaporation effects associated with surface rain gauges.

TRMM PR–based rain retrievals (Iguchi et al. 2000), designated as 2A25 in the TRMM data system, are applied as alternative rain products to validate the performance of 3B42 rain products. Wolff et al. (2005) demonstrate that 2A25 rain rates are in good agreement with ground radar measurements at TRMM ground validation sites. Yang et al. (2006a, 2008) and Yang and Smith (2008) show that 2A25 rain products are very comparable with TRMM microwave and microwave–precipitation radar combined rain products. However, the sampling issue associated with TRMM PR is obvious for studying small spatial systems because of its narrow swath. Both 3B42 and 2A25 rain estimates have relatively large errors at instantaneous and pixel resolution. This error decreases with the increase of spatial and temporal resolution. The version-6 TRMM rain products have an error of about 20% (3%) at the 14 mm h$^{-1}$ rain rate for the instantaneous 0.5° × 0.5° (monthly 2.5° × 2.5°) grid scale (Olson et al. 2006; Yang et al. 2006a). For June–August 1998–2001, ISCCP CTD, 3B42, and 2A25 rain datasets are used for this study.

The rectangular area (25°–40°N, 75°–105°E) is defined as the Tibetan domain for this study (Fig. 1). Any CS formed inside this domain is referred to as a Tibetan
CS. A larger area (Equator–40°N, 70°–140°E) is also used for studying the life cycle and propagation of summer Tibetan CSs.

b. Methodology

Because of the different spatial resolutions of ISCCP CTD variables and TRMM 3B42 rainfall, they must be reconstructed to have consistent analyses of rainfall and other properties. The 3B42 rain rates are selected to match the size of any tracked CS during its life cycle. Since both CTD variables and 3B42 rain rates are at the same 3-h interval temporal scale, a spatial matching procedure is applied. First, the size and coverage of any tracked CS at a designated time are determined by an ellipse equation based on its central position, eccentricity e, semimajor axis a, semiminor axis b, angle θ between the semimajor axis and the northward direction, and radius (r) of a circle with an equivalent area of this oval. The equations controlling the ellipse shape are \( a = r/\sqrt{e} \) and \( b = r\sqrt{e} \). Second, the 3B42 rain rates located in the oval area are matched to the pixels of this CS to have a spatial distribution of the CS surface rainfall. The CS center position is used for statistical analysis of CS initiation, tracking, and intensity. These procedures are followed for all tracked CSs.

3. Climatology of summer Tibetan CSs

a. Spatial distribution of CS frequency

A total of 643 CS life cycles originating from the Tibetan region are identified during the boreal summers of 1998–2001. The spatial distribution of these CS occurrences is virtually considered as climatology over the region. The distribution sheds light on the propagation of these CSs and their impacts on surface rainfall, which is particularly useful in rainfall prediction over the Tibetan and surrounding regions.

Figure 2a shows the spatial distribution of summer 1998–2001 Tibetan CS occurrences at 1.5° × 1.5° grid resolution identified with CTD. It is evident that the CSs originating from the Tibetan region moved over east China, South Asia, and the Bay of Bengal. Three salient features are clearly seen from Fig. 2a. First, two maximum centers are located inside the Tibetan region. This feature is expected since all of these CSs originated in the Tibetan region, which would lead to a high occurrence of CSs over the plateau. The dominant maximum in central-east/southern Tibet and the secondary center in central-southern Tibet have never been revealed in detail before. Second, the obvious broad eastward orientation of the relatively large amplitude of CS frequencies indicates the eastward movement of the CSs, which affects the rainfall over the Yangtze River basin, east China, and even the Korean Peninsula and Japan Sea. Third, the southward orientation of large amplitudes of CS frequencies shows a southward propagation of CS activities that mainly impact southwest China, the Indo-China Peninsula, and the Bay of Bengal. It is also evident that the southward movements are first concentrated in the southeast valley of the Tibetan Plateau, and then propagate with a southwest direction into the Bay of Bengal. This feature demonstrates that the Himalaya Mountains primarily block CSs direct southward propagation. In addition, there are rarely CS occurrences over the southeast coast of China where the west Pacific subtropical high persists in summer. This suggests that the propa-
The contribution of rainfall from summer Tibetan CSs to total precipitation is one of the important indices showing the impact of the Tibetan CSs on surface precipitation. Figure 2b presents a spatial distribution of rainfall contributions (%) from Tibetan CSs to total precipitation. It is evident that the maximum center with amplitudes >60% (up to a maximum of 76%) is located at the central-eastern area of the plateau. In addition, the Sichuan basin and the upper-middle segment of the Yangtze River basin have relatively large magnitudes (30%–40%). The contributions over other areas are about 10%–30%. Results suggest that rainfall from Tibetan CSs is an important contributor to total precipitation over the Tibetan Plateau and its surrounding regions. This is consistent with published results from Zhong et al. (1994), Yu (2001), and Zhang et al. (2001).

### b. Characteristics of Tibetan CSs

Five parameters extracted from the ISCCP datasets are utilized to present the characteristics of the summer Tibetan CSs: 1) $R_{cs}$—CS radius (size of CS); 2) $TB_{min}$—CS minimum cloud-top brightness temperature (intensity of convection); 3) $F_c$—percentage of strong convective fraction defined by the ratio of CC areas ($TB_{min} < 220$ K) to the CS coverage ($TB_{min} < 245$ K) (fraction of relatively stronger convection); 4) $R_{cc-max}$—maximum CC radius (size of the largest CC inside a CS); and 5) $C_{cc}$—CC count (number of CCs inside a CS).

Figure 3 displays the evolution of the averaged life cycles of categorized summer Tibetan CSs during 1998–
2001. These CSs are averaged at a 3-h temporal scale from their initiations over the plateau for different categories so that the evolution of the mean life cycles for the categorized CSs can be displayed. This approach demonstrates how the different categorized CSs evolve in terms of variability, intensity, size, propagation, and life span. It is obvious that the life span of CSs staying inside the Tibetan Plateau is only up to 66 h, while those CSs moving eastward and southward have a maximum life span of 192 and 288 h, respectively. \( \text{TB}_{\text{min}} \) decreases significantly and \( R_{cs} \) increases considerably in the first 3 h after the formation of a CS, while both \( R_{cc\text{-max}} \) and \( C_{cc} \) increase significantly. In general, \( \text{TB}_{\text{min}} \) of an eastward-moving CS continually decreases in the first 36-h life cycle, sustains the minimum cloud temperature for the next 36 h, and then gradually rises, while the variations of \( R_{cc} \), \( R_{cc\text{-max}} \), and \( C_{cc} \) are almost the opposite. However, the percentage of relatively stronger convective areas decreases significantly during the first 24 h and then gradually keeps decreasing. The southward-moving CSs have similar features to the eastward-moving CSs for the first 48 h; however, they continually intensify as \( \text{TB}_{\text{min}} \) decreases and \( R_{cc} / R_{cc\text{-max}} \) and \( C_{cc} / F_{c} \) increases. The similarities between the eastward- and southward-moving CSs during the first 2-day life cycle are possibly due to the fact that they are still strongly influenced by the thermodynamical forces on their way out of the Tibetan region, while dissimilarities are possibly caused by interactions with different environmental circulations during further development and propagation after moving out of the region. In addition, the southward-moving CS has a stronger intensity and a longer life cycle. The detailed physical mechanisms and associated environmental circulations for these two kinds of CSs are beyond the scope of this paper, but will be the subject of additional research.

It is also noticed that the curve for all CS cases is near the mean value of the three categorized CSs at the first 84 h of the CS life cycle, and the curve starts to bias toward the time series of the southward-moving CS thereafter. The following three reasons could possibly explain this feature: 1) the curve for all cases is not a weighted average of curves from the three categorized CSs so that CSs with more populations would have more impacts on the final mean curve; 2) the CSs moving southward are stronger and last longer than the other two categorized CSs; and 3) there is a relatively large frequency of the southward-moving CS populations whose life cycles are longer than 84 h (more discussion can be found in the description of Fig. 12). Therefore, the curve for all cases is not biased toward the southward-moving CSs at the early stage when contributions from CSs moving eastward and staying inside the Tibetan Plateau are equally important as the southward CSs; however, it shows a bias toward the southward-moving CSs after that early stage when influences from the southward-moving CSs increase. At the late stage of the CS life cycle, the all-CS curve is equivalent to the southward-moving CS curve because the contributions are entirely from southward-moving CSs.

The CSs show a strong diurnal variation during their life cycles (detailed analysis on diurnal behavior is in section 3c). The mean diurnal amplitudes of CSs staying inside the region are much larger than those moving out. This suggests that CSs staying inside the Tibetan region are mostly locally forced thermodynamic convection.

For comparison, the mean characteristics of summer Tibetan CSs at five different stages of life cycles are shown in Fig. 4. The five stages are defined as (i) initiation, (ii) \( \frac{1}{2} \) time between the initiation and the mature stage, (iii) mature, (iv) \( \frac{1}{2} \) time between mature and dissipation, and (v) dissipation. The initiation stage is
when a CS begins to be tracked. The mature stage is when the CS $TB_{\text{min}}$ is at its lowest magnitude; while the dissipation stage is the latest time of a tracked CS (similar features are found with defining the mature stage as $R_{cs}$ at its largest magnitude). It is evident that the CS $TB_{\text{min}}$ is higher at its initial and dissipation stages, as is expected. The CS $TB_{\text{min}}$ at the intermediate stages is between the lowest temperature at the mature stage and the initialization and dissipation stages. The corresponding change of CS size is the opposite of the variation of CS $TB_{\text{min}}$. The total number of CCs and the largest CC radius follow the same pattern. The fraction of relatively stronger convection area shows a different feature for the southward-moving CSs. Results demonstrate that the approach shown in Fig. 4 presents the composite features of CSs at five different stages of life cycles; however, it does not exhibit the evolution of the categorized Tibetan CSs.

The statistical mean and standard deviations of those variables shown in Fig. 3 are listed in Table 1. It is evident that the averaged $R_{cs}$ is about 340 km with $TB_{\text{min}}$ around 203 K. The relatively small $TB_{\text{min}}$ standard deviation shows that CS cloud-top height does not change dramatically, while the relatively large standard deviations of $R_{cs}$ (193 km) and $R_{cc\text{-max}}$ (79 km) indicate that CS sizes vary substantially. The mean CC count ($C_{cc}$) of 13 with a large standard deviation (14) demonstrates that the summer CS has a considerable CC variation at different stages of its life cycle. In addition, the relatively large $R_{cc\text{-max}}$ with respect to $R_{cs}$ shows that there are many small-sized CCs inside s. It suggests that one or more CCs favor more intensification than others inside a CS, indicating inhomogeneous development of CCs inside CSs. Therefore, these results hint at an asymmetrical development process during CS life cycles. The small $F_{c}$ suggests that the relatively deep convection activities are not active inside summer Tibetan CSs.

c. Diurnal cycle

Diurnal variation is one of the most important phenomena of atmospheric variables (e.g., Yang and Smith 2006, 2008; Yang et al. 2006b, 2008). The initiation of the summer Tibetan CS is influenced by the diurnal cycle of solar radiation and Tibetan thermodynamics. Case studies indicate that these CSs have a similar diurnal cycle with a frequency peak around 1800 local solar time (LST) (Zhu and Chen 2003); however, these case studies are not able to provide an accurate diurnal description of the Tibetan CS. Figure 5 exhibits an overall diurnal variation in the generation of summer Tibetan CSs in 1998–2001. A rapid increase of CS initiation around 1200 LST is clearly evident, with a maximum around 1500 LST. A total of 260 CSs occurred around 1500 LST, which accounts for about 40% of all CSs. A relatively weak secondary maximum is evident around 0600 LST. These results demonstrate that summer Tibetan CSs have a strong diurnal variability.

Figure 6 presents spatial distributions of the diurnal frequency of summer CS initiation over the Tibet Plateau. It is evident that the diurnal cycle of CS initiation has significant regional characteristics. The initiation of the summer CS increase when the sun rises, with a maximum around 1500 LST (when the surface heating is at peak). It also has a relatively large center over the central-eastern Tibet region in the midafternoon. A clear secondary maximum of the CS initiation is over the central-eastern region at 0600 LST. Thus, it appears that CSs favor the central-eastern region as their starting location.

Precipitation is one of the important indices in describing characteristics of the Tibetan CS. Figure 7a exhibits the diurnal variations of total precipitation and CS-only precipitation over the region for four summers (1998–2001). The total precipitation refers to rain from all raining systems, while the CS-only precipitation indicates rain only from Tibetan CSs. It illustrates that the summer total precipitation and CS-only-induced
rainfall have consistent prominent diurnal properties. The total precipitation has a dominant peak around 1800 LST and a secondary maximum around 0300 LST, while CS rain has a dominant peak around 0000–0300 LST and a secondary maximum around 1800 LST. The relatively larger late-night CS precipitation indicates in general that the intensity of the summer Tibetan CS is stronger in late night than in mid–late afternoon, suggesting that these CSs have a late-night intensification process. In addition, it is understandable that the diurnal amplitude of CS rainfall is much stronger than that of total precipitation. We have also shown in Fig. 2b that the CS-induced precipitation contribution to total precipitation is very important over the Tibetan Plateau and surrounding regions. Results illustrate that the CS-only-induced rainfall has a major impact on the diurnal variation of total precipitation.

Figure 7b illustrates the diurnal variations of rainfall from the categorized summer Tibetan CSs over their life cycles. The midafternoon and late-night rainfall peaks are obvious for CSs that do not move out of the region, while these CSs moving out of the region have a dominant late night peak. It is interesting to see that the CSs moving southward have a clear secondary rainfall peak. However, CSs propagating eastward do not have this feature; instead, a phase of the weak rainfall maximum is delayed a few hours.

These results clearly suggest possibly different
mechanisms for precipitation from summer Tibetan CSs. The midafternoon peak of CS rainfall suggests the forced surface heating, whereas the late night peak and the delayed afternoon maximum might be the combined outcome of four different mechanisms. The first is the so-called mobile terrain-forced precipitating system (MTFPS) mechanism summarized by Yang and Smith (2006). The mobility of the Tibet CS extends the rainfall maximum from afternoon into late afternoon–late night when the CS normally originates in the mid-

afternoon. The nighttime intensification process of the rainfall diurnal cycle described in the “static radiation–convection” and “dynamic radiation–convection” mechanisms proposed by Ramage (1971) and Gray and Jacobson (1977) could also explain that nighttime favors the development of the Tibetan CS. Huang et al. (1986) demonstrate that the heavy rainfall activities over southeast China are often accompanied by the late night peak of the summer low-level jet diurnal cycle. This could also link to the nighttime intensification process of the Tibetan CS propagating eastward and southward.

To independently check the 3B42 rainfall diurnal cycle, the TRMM PR–based rain datasets (2A25) over the Tibetan region for four summers (1998–2001) are also utilized in the diurnal analysis (Fig. 7c). It is clear that the total precipitation has a dominant afternoon peak at 1500–1800 LST with a secondary late night maximum at 0000–0600 LST. Further analysis of convective and stratiform rain shows that convective rain has only one prominent afternoon peak forced by the surface solar radiative heating, while stratiform rain has a relatively weak diurnal amplitude with a maximum at 0300–0600 LST and a secondary peak at 1500–1800 LST. The diurnal variations of the total and convective precipitation are very close, and the late-night peak of stratiform rain is primarily responsible for the weak late-evening maximum of the total precipitation. Yang and Smith (2008) and Yang et al. (2008) document in detail a global spatial distribution of the multiple modes of the rainfall diurnal cycle and the modulation of stratiform rain on the secondary peak. These facts demonstrate that the afternoon variability of rainfall in the Tibetan region is mainly controlled by convective activities forced by the surface solar heating, while stratiform rain leads to the secondary late-night peak. Comparisons of the total precipitation over the Tibetan Plateau in Figs. 7a,c show that diurnal amplitude is slightly smaller from 2A25 than from 3B42, and there is also a very small difference in the diurnal phase between them. The 2A25 slight underestimation is due to its physical assumptions in the algorithm, which have been discussed by many investigators (Wolff et al. 2005; Yang et al. 2006a; Kummerow et al. 2001). The very small difference in the diurnal phase might be caused by different instrument observations and sampling. However, the similarities between these two diurnal cycles are the dominant features, indicating a reliability of the TRMM 3B42 rain datasets applied in this study.

A phase shift is obvious between diurnal cycles of the summer Tibetan CS occurrences and the associated precipitation in Figs. 5, 7a. The phase of the afternoon

![Fig. 7. (a) Diurnal variability of the CS-only precipitation and total precipitation over the Tibetan region based on four-summer TRMM blended rain dataset (3B42). (b) Diurnal variation of precipitation from the categorized summer Tibetan CSs during their life cycles. (c) Diurnal cycle of the categorized rainfall from TRMM PR rain dataset (2A25) for same four summers; the all, convective, and stratiform precipitation from 2A25 are indicated by T, C, and S, respectively.](image-url)
rain maximum is about 3 h behind the CS occurrences because of the time needed for precipitation intensification processes associated with the CS developments.

A comparison of the summer Tibetan CS properties during daytime (0600–1800 LST) and nighttime (1800–0600 LST) is further conducted to explain the rainfall diurnal cycle discussed above. Table 2 lists the statistics of five variables representing the CS properties. They show that CSs have in general larger $R_{cs}$, $F_c$, $R_{cc-max}$, $C_{cc}$, and lower $TB_{min}$ with associated smaller standard deviations during nighttime than daytime, indicating that summer Tibetan CSs generally have a nighttime intensification process that could partially explain the dominant late-night peak of precipitation.

### Table 2. Differences of CS characteristics—means and std devs—during daytime and nighttime.

<table>
<thead>
<tr>
<th>Time</th>
<th>Parameter</th>
<th>$R_{cs}$ (km)</th>
<th>$TB_{min}$ (K)</th>
<th>$F_c$ (%)</th>
<th>$R_{cc-max}$ (km)</th>
<th>$C_{cc}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nighttime</td>
<td>Mean</td>
<td>359.8</td>
<td>201.6</td>
<td>24.0</td>
<td>127.0</td>
<td>13.4</td>
</tr>
<tr>
<td></td>
<td>Std dev</td>
<td>195.0</td>
<td>10.2</td>
<td>14.2</td>
<td>71.0</td>
<td>13.8</td>
</tr>
<tr>
<td>Daytime</td>
<td>Mean</td>
<td>318.2</td>
<td>205.2</td>
<td>20.9</td>
<td>111.0</td>
<td>12.7</td>
</tr>
<tr>
<td></td>
<td>Std dev</td>
<td>189.5</td>
<td>10.6</td>
<td>15.8</td>
<td>86.0</td>
<td>14.8</td>
</tr>
</tbody>
</table>

### 4. CS life cycle and development

#### a. Initiation of CSs

Figure 8 presents a spatial distribution of the summer Tibetan CS initiations. The prominent feature is the two local maximum centers near (30°N, 100°E) and (29°N, 92°E). This pattern is closely linked to activities of the south–north-oriented shear line over the Tibetan Plateau (Gao et al. 1984). The east maximum center is much stronger than the west center. This differs from findings by Jiang and Fan (2002) in which the west maximum center is stronger. Such a difference could be due to the different datasets involved in these two studies. The CS formed in the Tibetan region with a minimum radius of 55 km and life cycle of 3 h were included in Jiang and Fan (2002), while we focus on relatively larger CSs that have a minimum radius of 90 km and 6-h life spans. They also indicate that CSs that originated in the western center rarely move out of the plateau, which illustrates that these CSs have small sizes and short life spans, while those from the eastern center often propagate out of the region. Based on this study, CSs staying inside the Tibetan region have smaller and shorter life spans than those moving out; thus, more small-size and short-lived CSs in the Jiang and Fan...
(2002) study could shift the maximum center to the western region. In addition, there is a clear temporal variation in the regions where CSs formed (Fig. 9). The coverage and frequency of the summer Tibetan CS initiation increase from June to August. A slight northward shift of the frequency maximum centers is also visible. This is possibly due to the northward migration of maximum solar radiation.

Figure 10 shows the trajectories of all CSs that moved out of the Tibetan Plateau with eastward (orange lines) and southward (blue lines) propagations. It is evident that these mobile Tibetan CSs will generally impact rainfall in the south and east vicinity of the Tibetan Plateau. The eastward-moving CSs sometimes could affect rainfall over east China and even possibly the Korea–Japan region, while the southward-moving CSs could reach to the Bay of Bengal near the Thailand coast and occasionally to the Indian Ocean. It is rare to find any summer Tibetan CS reaching the southeast China coast area because of obstruction of the persistent summer west Pacific subtropical high.

The frequency statistics of the three categorized sum-
mer Tibet CSs are listed in Table 3, illustrating that 339 CSs (52.7% of total CS) did not move out of the Tibetan region, while 160 CSs (24.9%) moved out in an easterly direction and 144 CSs (22.4%) out toward the south. Therefore, more than one-half of the summer CSs die inside the region, while the remaining CSs virtually split between southerly and easterly propagations. There are slight annual variations of CS occurrences; however, these changes are small and not significant because of only four summer datasets. Nevertheless, it suggests a stable annual Tibet thermodynamical forcing.

b. Life cycle of CSs

The probability density function (PDF) and cumulative distribution function (CDF) are the commonly used parameters in illustrating statistical properties of meteorological variables. Figure 11 presents the (a) PDF and (b) CDF of the selected 643 summer Tibetan CSs in 1998–2001. It shows that both the PDF and CDF have a similar distribution pattern for each year, except for 1998 in which short-lived CSs dominated. Overall, the percentage of CS life spans in 12, 24, 36, and 48 h are about 23%, 26%, 16%, and 10%, respectively. The mean CDF curve indicates that almost 50% of the summer Tibetan CSs disappear in 24 h, with 75%, 85%, and 95% disappearing in 48, 60, and 108 h, respectively. Those features explain why over 50% of CSs did not move out of the region, because of their short life cycles. The occasional long life spans over 7 days might be because they merge into another system, or are simply due to uncertainties in the ISCCP CTD (Machado et al. 1998).

The PDF of the three categorized summer Tibetan CSs (Fig. 12a) show that CSs staying inside the plateau are dominated by short life cycles, suggesting that the localized system mainly depends on thermodynamical forcing. The east-propagating CSs have relatively more percentages than the south-propagating CSs for life spans of 48–72 h, indicating interactions with the surrounding atmospheric circulations. It is also apparent that there are more south-propagating CSs for life cycles longer than 84 h. Similar patterns exist for the PDF of precipitation associated with these three-
categorized CSs (Fig. 12b). The statistics of the six parameters describing the clear differences between CSs staying inside and moving out of the plateau are listed in Table 4. The mean $R_{cs}$, $TB_{min}$, $F_c$, $R_{cc-max}$, $C_{cc}$, and rain rate for CSs moving out of the Tibetan region is 429.7 km, 199 K, 25.9%, 150.7 km, 18.5, and 0.65 mm h$^{-1}$, while those staying in the region are 291.5 km, 204.7 K, 21.4%, 100.2 km, 9.7, and 0.45 mm h$^{-1}$, respectively. They show a 50% increase for $R_{cs}$ and $R_{cc-max}$, 5% rise in relatively stronger convection, and 100% more $C_{cc}$ as well as a 45% increase of rain intensity, from CSs staying in the region to those moving out of the area. Results illustrate that CSs moving out of the Tibetan region have more rainfall and longer life cycles when compared with those staying inside the plateau. These also suggest influences of easterly-moving CSs on the upper-middle segments of the Yangtze River basin, and potential impacts on faraway regions from the southerly-propagating CSs. These results are consistent with the findings by Yu and Gao (2006) that 93% of the Tibetan vortexes moving out of the Tibetan region produce more rainfall than those staying inside the region. Those CSs moving out of the plateau appear to have favorable environmental circulations that extend their life cycles. It also indicates that the surrounding atmospheric conditions play an important role in the development of the summer Tibetan CS.

5. Discussion and conclusions

All CSs originating from the Tibetan Plateau defined by the ISCCP deep convection database along with associated TRMM precipitation datasets for four summers (June–August 1998–2001) have been analyzed to characterize their occurrence frequency, spatial distribution, development, life cycle, track, and precipitation.

There are three types of summer CSs that have their initiation over the Tibetan Plateau. The first one does not move out of the plateau and dominates the CS population with 53% of frequency. The second category, counted at 25% of the CS population, propagates eastward out of the Tibetan Plateau and impacts primarily precipitation over the upper-middle segment of the Yangtze River basin. The third type is CSs moving southward out of the region, which accounts for about 22% of all Tibetan CSs. This type of CS produces heavy rainfall in the southwest of China, Thailand, and even to the east of the Bay of Bengal. The CS rainfall contribution to total precipitation is up to 76% over the

<table>
<thead>
<tr>
<th>CS category</th>
<th>Parameter</th>
<th>$R_{cs}$ (km)</th>
<th>$TB_{min}$ (K)</th>
<th>$F_c$ (%)</th>
<th>$R_{cc-max}$ (km)</th>
<th>$C_{cc}$</th>
<th>Rain rate (mm h$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Move out</td>
<td>Mean</td>
<td>429.7</td>
<td>199.0</td>
<td>25.9</td>
<td>150.7</td>
<td>18.5</td>
<td>0.65</td>
</tr>
<tr>
<td></td>
<td>Std dev</td>
<td>224.6</td>
<td>9.5</td>
<td>13.6</td>
<td>82.9</td>
<td>18.5</td>
<td>0.44</td>
</tr>
<tr>
<td>Not out</td>
<td>Mean</td>
<td>291.5</td>
<td>204.7</td>
<td>21.4</td>
<td>100.2</td>
<td>9.7</td>
<td>0.45</td>
</tr>
<tr>
<td></td>
<td>Std dev</td>
<td>142.2</td>
<td>9.9</td>
<td>15.3</td>
<td>60.3</td>
<td>9.4</td>
<td>0.38</td>
</tr>
</tbody>
</table>

Fig. 12. (a) PDF of summer Tibetan CS life cycles for those moving easterly and southerly out of the region, and those staying inside. (b) As in (a), but for associated surface rainfall.
central-eastern area of the Tibetan Plateau, and 30%-40% over the east-southern adjacent regions, showing important impacts of summer Tibetan CSs on regional total precipitation.

The two maximum initiation centers of summer Tibetan CSs are located to the south of 33°N, one over the Yaluzangbu River valley to the west of 95°E and another over the Hengduan Mountain valley to the east of 95°E. The east center is much stronger, and moves slightly northward from June to August because of the northward migration of the sun.

The summer Tibetan CS has a mean life span of about 36 h. Approximately 85% of the Tibetan CSs disappear within 60 h of their initiations. The southward-moving CS has the longest life cycle, while those CSs staying inside the Tibetan Plateau have the shortest life cycle. The relatively stronger convection area is only about 22% of the CS coverage. Imbedded CCs clearly show the CS initiation–die processes and have an inhomogeneous distribution. Results also demonstrate that the more intense a Tibetan CS is, the farther away the maximum CC is from its center, indicating an asymmetric development process during the CS life cycle. In addition, CSs staying inside the Tibetan Plateau are generally smaller, have shorter life spans, and produce less rainfall than those moving out of this region. The differences between these two types of CSs suggest that the former is mainly forced by the surface heating while the latter is due to interactions with the surrounding atmospheric circulations during their life cycles. The detailed physical mechanisms behind these CSs are the subject of a follow-up study.

Initiation of CSs occur mainly in the afternoon with a maximum around 1500 LST. The maximum frequency in midafternoon is about 6 times the minimum frequency at midnight, demonstrating that surface solar heating is the primary force for initiation of the summer Tibetan CS. The CS development and associated rainfall also have obvious diurnal cycles. It appears that stratiform rainfall has a distinct late night peak and midafternoon maximum while convective rainfall shows a major midafternoon peak, indicating possibly multiple mechanisms controlling the diurnal cycle of rainfall over the Tibetan Plateau. The afternoon rainfall peak corresponds to the convection forced by the surface solar radiative heating. The late night maximum is linked to “static radiation–convection” and/or “dynamic radiation–convection” mechanisms as well as the low-level jet diurnal cycle. In addition, CSs initiated in midafternoon at one location could bring delayed mid-afternoon–late night maximum rainfall to another area. This process is summarized as the MTFPS mechanism by Yang and Smith (2006). The prominent peak of the CS precipitation in late night also suggests a nighttime intensification process during the development of the summer Tibetan CSs, indicating that the mechanism for the CS development is different from its initiation.

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