Evolution of Warm Season Intense Rainfall in Yaan Against a Cold-Anomaly Background

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Abstract

This study investigates the rainfall characteristics during intense rainfall over Yaan against a cold-anomaly background, aiming to refine the understanding of different kinds of rainfall events across complex terrain. Hourly rain gauge records, ERA5 reanalysis data and the black body temperature of cloud tops derived from FY-2E were used. The results show that against a cold-anomaly background, the regional rainfall events (RREs) in Yaan exhibit west-to-east propagation, which is different from the north-to-south evolution of warm RREs. The middle and upper troposphere is dominated by a negative geopotential height anomaly corresponding to the cold anomaly. The cyclonic circulation at the higher level associated with the negative geopotential height anomaly bends the high-level jet to the south, forming a divergent zone over the Tibetan Plateau (TP) and guiding mid-level systems to move eastward. The cyclonic circulation at the mid-level produces a wind shear zone over the TP, generating anomalous vorticity that continuously moves eastward and develops to influence the rainfall over Yaan. The cold Yaan RREs are closely related to the TP low-pressure systems (both vortex and shearline). The anomalous vorticity over the TP can influence the local vortex over the eastern periphery of the TP at a distance mainly by the horizontal advection of anomalous vorticity by the mean flow and then enhance the local vortex mainly by anomalous convergence when it moves near Yaan.

1 Introduction

Precipitation is one of the most important subjects in both weather forecasts and climate projections. However, the performance of precipitation simulations in numerical weather and climate models is not satisfactory, especially over complex terrain (Meehl and Arblaster, 1998; Rucong et al., 2000; Carbone et al., 2002; Phillips and Gleckler, 2006; Ying and Chong-Hai, 2012; Gao et al., 2015). In addition to problems within the numerical models themselves, the intricacy of precipitation mechanisms over complex terrain is one of the main causes of the relatively poor performance. Categorizing different kinds of rainfall and studying the evolution characteristics of the rainfall process through observations may help in both understanding the precipitation mechanism and evaluating the model.

Connecting the Tibetan Plateau (TP) in the west and the Sichuan Basin in the east, the eastern periphery of the Tibetan Plateau (EPTP) is a typical region with complex terrain (Fig. 1). The rainfall over the EPTP, characterized by that at the well-known “rainy city” Yaan, was modulated by different scales of topography (Li et al., 2010a). For example, in addition to the immense relief of the TP (Jin et al., 2013), the local terrain of Yaan, which is funnel-shaped and opens to the east, could also favor convergence and precipitation when the easterlies blow toward the EPTP (Xi, 1992; Peng et al., 1994; Zeng et al., 1995). Moreover, varied circulation systems exert an influence on rainfall over the EPTP. For example, the vortex originating from the TP (known as the TP vortex) and that in the Sichuan Basin (Southwest vortex) could both result in EPTP rainfall and sometimes interact with each other (Li et al., 2017a; Li et al., 2020). Other circulation systems, such as synoptic cold fronts, the low-level jet, the western Pacific subtropical high, and even typhoons, also have impacts on the EPTP rainfall characteristics, such as intensity and diurnal variations (Wang, 1977; Peng et al., 2008). The comprehensive influence of topography and circulation

systems has made the EPTP a typically biased region in precipitation simulations (Li et al., 2015a), and studies that classify the rainfall characteristics and mechanisms over the EPTP are therefore urgent.

With the help of hourly rain gauge records, subdaily fine-scale rainfall characteristics have been comprehensively revealed within inland China in recent decades, including diurnal variations, the intensity-frequency structure of overall rainfall, and the duration and asymmetry of rainfall events (Yu et al., 2007a; Yu et al., 2007b; Yu et al., 2013; Li and Yu, 2014; Li, 2018). When regional rainfall events (RREs) are defined, the evolution of the rainfall influencing a specific region can be determined, which enables understanding of the mechanism and identification of precursors (Sun et al., 2015; Yu et al., 2015; Sun et al., 2018; Yuan et al., 2018). For example, Sun et al. (2018) distinguished two kinds of rainfall events with different propagation processes: when upper tropospheric warm anomalies dominate North China, rainfall tends to remain in the western mountain region, and when cold anomalies dominate North China, rainfall propagates to the eastern plain. Such results are valuable for rainfall forecasts along the precipitation pathway. On the EPTP, the fine-scale rainfall characteristics are also under ongoing study. For example, Chen et al. (2017) investigated distinct nocturnal rainfall and related convection activity over the EPTP and found that rainfall events in this region are characterized by frequent moderate rainfall with long durations, which differs from those of both the eastern TP and the western Sichuan Basin. Their results show evidence of the close relationship between the RREs with early TP rainfall and the eastward propagation of the convective systems that develop over the TP in the late afternoon. The evolution of warm season intense RREs in Yaan and the associated three-dimensional circulation have also been analyzed (Hu et al., 2020). The composite results showed that warm anomalies were present in the upper troposphere before and during the RREs and were associated with north-to-south propagation of rainfall. However, not all RREs are controlled by a warm anomaly in the upper troposphere. Whether a cold anomaly exists before an RRE and how the evolution proceeds against such a background remain open questions worth further investigation.

This study aims to answer these questions by analyzing first the rainfall characteristics and then the evolution of the circulation. The data and method used are described in Section 2, followed by the rainfall characteristics in Section 3 and the evolution of the corresponding circulation in Section 4. Discussion and conclusions are given in Sections 5 and 6, respectively.

2 Data And Methods

2.1 Observational and reanalysis data

The quality-controlled hourly rain gauge records provided by the National Meteorological Information Center of the China Meteorology Administration were used here, covering the Sichuan Basin and the time range of warm seasons (May-October) from 1986-2020 (Fig. 1). Each of the 1102 stations had valid records for more than 60% of the total hours in the entire time range. A detailed description of the dataset was documented in Zhang et al. (2016). The latest ERA5 reanalysis data (spatial resolution $0.25^{\circ} \times 0.25^{\circ}$) for the same time range were also used in this study, which were provided by the European Centre for

Medium-range Weather Forecasts and developed through the Copernicus Climate Change Service (Hersbach et al., 2019). Following Hu et al. (2020), the anomalous circulation was calculated by removing the multiyear mean of the corresponding time steps for 11 days around the analyzed time step to avoid intraseasonal and diurnal variability.

2.2 Regional rainfall events

To study the evolution of rainfall events that affect a limited area around Yaan, we applied the method of RREs suggested by Yu et al. (2015). The RRE method is well suited to study the spatial-temporal evolution characteristics of rainfall processes in a specific region. For every hour t , the maximum hourly intensity $P_{x,t} = \max(P_{i,t}) (i = 1, N)$ is determined within N stations in a selected key region, where $P_{i,t}$ is the measured rainfall at station i for hour t (here, x refers to the maximum). The time series of $P_{x,t}$ thus represents the precipitation intensity of the key region, and an RRE is then defined as a period of consequent time series above the threshold (0.1 mm h^{-1} here) with at most a 1-h intermittency, following Yu et al. (2015). Each RRE has its starting, peak and ending times, as well as its spatial features. To quantitatively investigate the spatial features of RREs, the regional rainfall coefficient (RRC) is defined following Yu et al. (2015):

$$RRC = \left[2 \times \frac{P_{mean}}{P_{max} + P_{min}} \right] \times \left[\frac{N_a - 1}{N - 1} \right]$$

where $N(N_a)$ is the total number of stations (rainy stations) in a target region and P_{mean} , P_{max} and P_{min} are the area mean, maximum and minimum rainfall amounts, respectively, observed at all rainy stations. The RRC is thus a comprehensive measure of both the spatial scale and the uniformity of precipitation.

Similar to Hu et al. (2020), we selected intense RREs based on the following criteria: (1) the process maximum intensity of the RRE was greater than 10 mm h^{-1} , (2) the number of hours from the start to the peak of the RRE was less than 8 h, and (3) the RRC was greater than 0.4. In total, 619 intense RREs were selected based on these criteria. Moreover, to preclude influence between intense RREs, we focused on only the 459 RREs for which no other intense RREs occurred in the 12 h prior to the start and after the end of the event. The empirical orthogonal function (EOF) was applied to the temperature anomaly field at 300 hPa for 6 h before an event, as the high-level temperature anomaly is a significant characteristic of the Yaan RREs (Hu et al., 2020). The first mode of the EOF accounts for approximately 44% of the total variance, showing a monopole pattern centered over northern China (Fig. 2a). This monopole pattern with a large explanatory variance indicates that the temperature anomaly varies widely in this region. According to the principal component of the first EOF mode (Fig. 2b), RREs with cold anomalies are conveniently separated from those with warm anomalies. The RREs are hence categorized as warm-anomaly RREs if their principal component is less than $-\sigma$ or as cold-anomaly RREs if their principal component is greater than $+\sigma$, where σ equals the standard deviation of the principal component. In

final, 86 cold-anomaly RREs and 67 warm-anomaly RREs are selected. Fig. 2c and d present the composite 300-hPa temperature anomalies of cold- and warm-anomaly RREs, respectively. Cold and warm anomalies are clearly separated, and the center of a warm anomaly is larger and heavier than that of a cold anomaly. As the features of composite results for warm anomalies are shown in Hu et al. (2020), cold-anomaly RREs are the main focus of this study.

2.3 Decomposition of anomalous vorticity tendency

Because Yaan rainfall was the main focus of this study, the tendency of anomalous vorticity over the EPTP was diagnosed to investigate how local systems developed over Yaan. The tendency of anomalous vorticity in isobaric coordinates can be expressed as follows:

$$\begin{aligned} \frac{\partial \zeta'}{\partial t} = & - \left(\bar{u} \frac{\partial \zeta'}{\partial x} + \bar{v} \frac{\partial \zeta'}{\partial y} \right) - \left(u' \frac{\partial \bar{\zeta}}{\partial x} + v' \frac{\partial \bar{\zeta}}{\partial y} + v' \frac{\partial \bar{f}}{\partial y} \right) - \left(\bar{\omega} \frac{\partial \zeta'}{\partial p} + \omega' \frac{\partial \bar{\zeta}}{\partial p} \right) \\ & - \left(\bar{\zeta} + f \right) \left(\frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} \right) - \zeta' \left(\frac{\partial \bar{u}}{\partial x} + \frac{\partial \bar{v}}{\partial y} \right) \\ & + \frac{\partial \bar{\omega}}{\partial y} \frac{\partial u'}{\partial p} - \frac{\partial \bar{\omega}}{\partial x} \frac{\partial v'}{\partial p} + \frac{\partial \omega'}{\partial y} \frac{\partial \bar{u}}{\partial p} - \frac{\partial \omega'}{\partial x} \frac{\partial \bar{v}}{\partial p} + Res \# (1.) \end{aligned}$$

in which anomaly $X' = X - \bar{X}$. The variable names are those in customary use in meteorology. *Res* contains a friction term and small second-order quantities.

3 Rainfall Characteristics

A comparison of the rainfall characteristics between cold and warm RREs was first conducted. Fig. 3 presents the joint probability density functions (PDFs) of event numbers between diurnal cycles of start times and durations, maximum intensity and the RRC of cold and warm events. In both cold and warm events, a dominant nocturnal start time can be found, consistent with the robust nighttime peak in rainfall there (Lee, 1942). The difference is that most warm events start in the early half of the night before 01:00 BJT (Fig. 3d), while the spread of the nocturnal peak in cold events is larger, with a portion of cold events starting in the early morning (Fig. 3a). Cold events with durations shorter than 10 h can start in a broad time range from approximately 16:00 to 07:00 BJT the next day, and longer events with durations greater than 15 h mostly start from 16:00 to 22:00 BJT (Fig. 3a), suggesting a longer-earlier relationship. The same relationship can also be found in warm events (Fig. 3d). The distributions of the maximum intensity of all RREs are shown in Fig. 3b and e for cold and warm events, respectively. For cold events (Fig. 3b), the frequency of the maximum intensity decreases dramatically as the intensity increases. Many cold events have a maximum intensity ranging from 10-20 mm/h. Meanwhile, most of the cold events have an RRC larger than 0.7 (Fig. 3c). For warm events, the situation is different. The frequency of

the maximum intensity varies from 10 to 40 mm/h (Fig. 3e), and considerable numbers of these events have RRC lower than 0.7 (Fig. 3f), suggesting a higher frequency of events with stronger convection and smaller rainfall area during warm anomalies. In contrast, a cold background seems to favor more systematic although lighter rainfall.

The rainfall evolution characteristics of cold events are shown in Fig. 4 as the hourly rainfall frequency. Unlike the north-to-south propagation of warm anomalies documented in Hu et al. (2020), rainfall presents a west-to-east evolution in cold events. From 12 to 3 h prior to the start of cold RREs, the stations on the eastern part of the Tibetan Plateau show a larger rainfall frequency, meaning that rainfall has a higher probability of originating there than elsewhere. Then, the rainfall moves to the key region, intensifying and expanding its spatial range, corresponding well with the large RRCs shown in Fig. 3c. Approximately 9 h after the start time, the rainfall weakens and moves farther east. The clear propagation of the rainfall frequency depicted in Fig. 4 implies that the rainfall at Yaan on the EPTP may be traced back to the plateau, while the specific connection needs to be clarified by examining the circulation systems in detail.

4 Corresponding Circulations

4.1 The circulation structure before cold RREs occur

Figure 5 gives the circulation at different isobaric levels from the upper to lower troposphere 3 h before the start time of cold RREs. Because of the cold anomaly at 300 hPa centered at approximately 113°E, 35°N, there is a negative geopotential height anomaly above 200 hPa according to the thermal wind equilibrium and hydrostatic balance (Fig. 5a). The cyclonic circulation reduces the western wind speed to the north and enhances the western wind to the south, bending the jet core axis toward the south. Immediately south of the high-level jet, a large divergent zone forms over the eastern TP and near the EPTP where rainfall is about to occur. The cold anomaly stretches down to 500 hPa, although it is not as cool as that at the upper level (Fig. 5b). In addition, the shapes of the cold region are different. At the upper level, the cold core extends from northwest to southeast, and the northwestern end is narrower. Accordingly, the curvature there is larger. At 500 hPa, however, the cold region extends from west to east with a wedge-like cold region over the eastern TP. Accordingly, the anomalous winds in this region show a large curvature, and a region of high vorticity can be found there. The EPTP is now immediately before an anomalous trough, and the Sichuan Basin is controlled by an anomalous ridge. The cold anomaly decreases as the level decreases. At 850 hPa (Fig. 5c), a warm anomaly appears over the Sichuan Basin. The warm anomaly is more significant along the southwestern part of the basin, where the southeasterly winds blow toward the EPTP, supplying moisture for rainfall. To clarify the vertical structure of the circulation, a cross-section along 90°-110°E averaged over 30°-33°N (black box in Fig. 5c) is shown in Fig. 5d. The cold anomaly extends from approximately 150 hPa down to 750 hPa, centered at 300 hPa. Notably, in addition to the broad cold anomaly in the east, a region on the surface of the TP has an anomalous temperature less than -1.5 K, corresponding to the wedge-like cold anomaly over the TP shown in Fig. 5b. A large divergent zone covers the eastern TP and the EPTP at the upper level, and two

regions with large vorticity can be discerned, with one over the TP extending from 500 hPa to 250 hPa and the other over the EPTP linked to the topography.

4.2 Evolution of the circulations

We next examined the evolution of the circulations corresponding to the rainfall processes.

The evolution of upper-level circulations at 6-h intervals is shown in Fig. 6 from 12 h before to 6 h after the start time of the cold RREs. At -12 h, a circular-shaped cold anomaly is present northeast of the TP. The high-level jet bends to the south in association with the cold anomaly. Divergence develops on the right side of the jet core over the TP at approximately 95°E, 33°N. This result is consistent with the situation of the long-duration TP vortex depicted by Yu and Gao (2018). In later hours, the circular-shaped cold center tends to stretch in the northwest-southeast direction and to weaken. The eccentricity of the anomalous geopotential height increases, with the high-level jet extending progressively to the east and connecting with the eastern part 6 h after the start time. In this process, the divergent zone propagates from the TP at -12 h to the EPTP when rainfall starts in the key region and then farther east after the beginning of rainfall.

The evolution of the 500-hPa circulation is shown in Fig. 7. At -12 h, the main part of the cold anomaly is located outside the TP centered at 110°E, 35°N, immediately below that at 300 hPa. The northerlies on the western edge of the cyclonic circulation associated with the cold anomaly blow toward the TP, and a wind shear line forms along 33°N at approximately 92°E. At -6 h, although the center of the cold anomaly decreases, a cold tongue appears over the TP, and the wind shear intensifies. The vorticity associated with the wind shear line moves eastward in the next few timesteps. The dominant mid-level systems change to an anomalous trough when the RREs start from a ridge 12 h prior.

At 850 hPa (Fig. 8), most parts of China are still controlled by cold anomalies except the Sichuan Basin and some areas north of the TP. Northerly winds dominate eastern and southern China at -12 h, and anomalous anticyclonic circulation develops in the Sichuan Basin, which is not favorable for precipitation at this time. In the next few hours, the southerlies over the southern edge of the Sichuan Basin are enhanced as the warm anomaly intensifies. As the mid-level trough moves near the basin, the wind follows a cyclonic path and turns toward the EPTP, favoring rainfall there.

The evolution of the vertical structure of the circulation is given in Fig. 9. At -12 h, the center of the cold anomaly is located at approximately 110°E and 300 hPa, and the -1.5 K contour is separated from the TP surface. Weak vorticity can be found over the TP at 95°E and 400 hPa and over the EPTP near the ground surface. Upward motion is associated with the vorticity over the TP, but the air mass diverges at approximately 200 hPa; the winds turn westerly at the upper level and then descend over the EPTP and the Sichuan Basin, making the low-level air divergent there. At -6 h, the cold center moves away from the eastern boundary. The vorticity and the high-level divergence over the TP both develop distinctly in intensity and width. Upward motion persists over the whole extent of the TP, with the maximum vertical velocity in the coupled zone between the mid-level vorticity and the high-level divergence at approximately

97°E. The vorticity over the EPTP also increases in intensity but is somehow still suppressed near the surface. Weak ascending motion appears over the EPTP at this timestep. When the RRE starts, the cold anomaly moves farther east associated with the high-level divergence and vorticity over the TP. The high-level divergence occurs immediately over the EPTP, and the vorticity from the TP merges with the vorticity over the EPTP. Consequently, large vertical motion develops over the EPTP, favoring rainfall there. Notably, the -1.5 K temperature contour is attached to the TP surface at this time. In later hours, systems continue to propagate eastward. The vorticity over the EPTP develops to mid-level from the surface as the systems from the TP move in.

To further understand the evolution of the convection associated with the system propagation and rainfall process, the convection index is calculated and presented in Fig. 10 with the mid-level vorticity and the rainfall frequency. Although the centers of the three variables are not coincident with each other, they all show west-to-east propagation. Corresponding to the vorticity, the convection originating from the TP at approximately 98°E and 6 h reaches its maximum at approximately 101°E and 2 h and then propagates farther east. The rainfall frequency then reaches its peak at 103°E and 6 h, in a region that is downstream of the center of convection. Fig. 10 indicates that the cold Yaan RREs have close connections with the systems and convections upstream on the TP.

5 Discussion

In the above sections, we analyzed the evolution characteristics of rainfall during the Yaan RREs against a cold-anomaly background, as well as the corresponding circulations. The rainfall is characterized by west-to-east propagation from the TP to the basin, and the circulation is characterized by cold anomalies, a strong and south-directed high-level jet and anomalous wind shearlines originating from the TP. The results are consistent with those from previous research on the TP vortex and shearlines. When studying the long-duration TP vortex moving off the plateau, Yu and Gao (2018) also found that a strong high-level jet appeared, which could transfer the 200-hPa high potential vorticity to the vortex when the jet extended to the east and south, as they pointed out. Based on atmospheric motion equations and a divergence equation, Du et al. (2020) proved that the transverse shearline over the TP is an important background field of the TP vortex. Consistency with previous studies suggests that the cold Yaan RREs could be related to the TP low-pressure systems in the primitive field, which have been extensively studied. Therefore, the numbers of TP low-pressure systems related to cold RREs were counted from the almanacs of the TP vortex and shearlines (Li et al., 1998, 1999, 2000, 2001, 2002, 2003, 2004, 2005, 2006, 2007, 2008, 2009, 2010b, 2011, 2012, 2013, 2014, 2015b, 2016, 2017b, 2018). The TP low-pressure systems related to cold RREs were selected based on two criteria: (1) the TP system occurred in the time range from 2 days before the start time to the end time of the cold RREs, and (2) the minimum distance between the path of the TP system and Yaan station was less than 3°. These criteria were set to ensure that the TP system was sufficiently close to Yaan RREs in both time and space. For comparison, TP low-pressure systems related to warm RREs were also counted. The paths of the TP systems are shown in Fig. 11. Coincidentally, there were 44 cold- and 44 warm-anomaly RREs from 1998 to 2018, of which 18 and 9, respectively, had corresponding RRE-related TP systems. In total, there were 27 TP low-pressure

systems related to cold RREs and 9 systems related to warm RREs. Compared with warm-anomaly RREs, cold-anomaly Yaan RREs are clearly more closely related to TP systems.

The TP low-pressure systems, regardless of whether they move off the TP, can exert influence on the rainfall to the east of the plateau and on the southwestern vortex in the basin (Li et al., 2020). To investigate how the local vortex over the EPTP develops under the influence of upstream TP systems, equation (1) is applied to the region 30-33°N, 102-104°E over the EPTP. Fig. 12 shows the average results in the area for each component of the equation. The anomalous vorticity develops mainly before the beginning of an RRE (Fig. 12a). Prior to -4 h, the vorticity tendency is contributed mainly by the friction term below 500 hPa (Fig. 12j), which is consistent with the near-surface vorticity observed in Fig. 9a. Above 500 hPa, the horizontal advection of the anomalous vorticity caused by the averaged flow

$-\left(\bar{u}\frac{\partial\zeta'}{\partial x} + \bar{v}\frac{\partial\zeta'}{\partial y}\right)$ is the largest term contributing to the tendency (Fig. 12b). As westerlies prevail

above 500 hPa in the mean state, the anomalous low-pressure systems over the TP could influence the EPTP vorticity from a distance by vorticity advection. After -4 h, the anomalous divergence-related term

$-\left(\bar{\zeta} + f\right)\left(\frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y}\right)$ dominates the vorticity tendency (Fig. 12f). This term increases as the vorticity

from the TP moves near the EPTP, causing convergence and convection and consequently enhancing the local vorticity. Thus, two pathways may exist by which an anomalous low-pressure system from the TP can influence the vorticity over the EPTP. When the TP system is on the plateau, it can affect the local vortex over the EPTP by the horizontal advection of anomalous vorticity with the mean flow, and when the anomalous TP system moves close to the EPTP, direct anomalous convergence can enhance the vorticity.

6 Conclusions

Using hourly rain gauge records, the rainfall characteristics of Yaan regional rainfall events against a cold-anomaly background were investigated. The corresponding evolution of the circulation and convection was then analyzed by using the ERA5 reanalysis and FY-2E TBB datasets. The main conclusions are summarized as follows:

1. Against a cold-anomaly background, the rainfall of Yaan RREs exhibits west-to-east propagation, which is different from the north-to-south evolution for warm RREs. Before the beginning of cold Yaan RREs, the rainfall frequency at stations over the eastern TP is the largest, suggesting that rainfall tends to originate from there. Then, the rainfall moves farther east, influencing Yaan and expanding in the basin.
2. The middle and upper troposphere is dominated by a negative geopotential height anomaly corresponding to the cold anomaly. The cyclonic circulation at the higher level associated with the negative geopotential height anomaly bends the high-level jet to the south, forming a divergent zone

over the TP and guiding mid-level systems to move eastward. The cyclonic circulation in the mid-level produces a wind shear zone over the TP, generating anomalous vorticity that continuously moves eastward and develops to influence the rainfall over Yaan.

3. The cold Yaan RREs are closely related to the TP low-pressure systems (both vortex and shearline). When a low-pressure system is generated over the TP, the local vortex over the EPTP can be influenced mainly by the horizontal advection of anomalous vorticity by the mean flow. When the low-pressure system moves near Yaan, the local vorticity is enhanced mainly by anomalous convergence associated with the TP system.

Declarations

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Data Availability: The hourly rain-gauge records and the black body temperature of cloud tops derived from FY-2E analyzed during the current study are available at <http://data.cma.cn>. The ERA5 reanalysis data used in the present study is available at <https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-pressure-levels?tab=form>.

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Figures

Figure 1

Stations used in this study and the topographic distribution. There are 1102 stations in total, each of which has valid records for more than 60% of the total hours in the warm season (May to October) from 1986 to 2020. The red dots represent the 9 key stations around Yaan. The color shading represents surface elevation (unit: m).

Figure 2

The first mode of the EOF of the temperature anomaly 6 h prior to the start of the warm season intense Yaan RREs from 1986 to 2020 (a) and corresponding principal component (b). RREs whose PC coefficient is less than minus one standard deviation are categorized as cold RREs, and those larger than one standard deviation are categorized as warm RREs. The composite results for the 300-hPa temperature anomalies (unit: K) of cold and warm RREs are shown in (c) and (d), respectively.

Figure 3

The joint probability density functions of the number of RREs with diurnal cycles of start times and the durations (a, d), maximum intensities (b, e) and RRCs (c, f) of cold and warm events, respectively.

Figure 4

Spatiotemporal evolution of the precipitation frequency (unit: %) during the 12 h before and after the start of the cold Yaan RREs.

Figure 5

The three-dimensional circulation structure in the troposphere at 3 h prior to the start of the cold Yaan RREs. The black and red lines in (a) represent the 200-hPa geopotential height anomaly (unit: gpm) and the divergence anomaly (unit: 10^{-5} s^{-1} , the contour starts from 0.75 to 1.75 with interval 0.5), and the cyan line is the 30 m/s contour of the wind speed in the primitive field. The color shading represents the 300-hPa temperature anomaly (unit: K). The white line in (b) represents the vorticity anomaly (unit: 10^{-5} s^{-1} , the contour starts from 1.0 to 2.5 with interval 0.75) at 500 hPa, and the vector and color shading are the anomalous horizontal wind (unit: m s^{-1}) and temperature at 500 hPa, respectively. The vector and

color shading in (c) are the anomalous horizontal wind and temperature at 850 hPa, respectively. (d) gives the cross-section for the black box shown in (c). The stippling indicates places where the temperature field is significant at the 99% level of Student's t test, and only the significant winds are drawn in black.

Figure 6

Circulation evolution in the upper troposphere from 12 h before to 6 h after the start of the cold Yaan RREs. The black and red lines represent the 200-hPa geopotential height anomaly (unit: gpm) and the divergence anomaly (10^{-5} s^{-1} , the contour starts from 0.75 to 1.75 with interval 0.5), and the cyan line is the 30 m/s contour of the wind speed in the primitive field. The color shading represents the 300-hPa

temperature anomaly (unit: K). The stippling indicates places where the temperature field is significant at the 99% level of Student's t test.

Figure 7

Circulation evolution in the middle troposphere from 12 h before to 6 h after the start of the cold Yaan RREs. The white line represents the vorticity anomaly (10^{-5} s^{-1} , the contour starts from 1.0 to 2.5 with interval 0.75) at 500 hPa, and the vector and color shading are the anomalous horizontal wind (unit: m s^{-1}) and temperature (unit: K) at 500 hPa, respectively. The stippling indicates places where the temperature field is significant at the 99% level of Student's t test, and only the significant horizontal winds are drawn.

Figure 8

Circulation evolution in the lower troposphere from 12 h before to 6 h after the start of the cold Yaan RREs. The vector and color shading are the anomalous horizontal wind (unit: m s⁻¹) and temperature (unit: K) at 850 hPa, respectively. The stippling indicates places where the temperature field is significant at the 99% level of Student's t test, and only the significant horizontal winds are drawn in black.

Figure 9

The evolution of the vertical structure of the circulation along the black box shown in Fig. 5c. The red (white) contours represent divergence (vorticity) anomalies (unit: 10^{-5} s^{-1}). The color shading (vector) presents the temperature (wind) anomalies. The unit of the vertical-component wind is -102 Pa s^{-1} , and that of horizontal wind is m s^{-1} . The stippling indicates places where the temperature field is significant at the 99% level of Student's t test, and only the significant winds are drawn in black.

Figure 10

Hovmueller plots of relative vorticity at 500 hPa (a, unit: 10^{-5} s^{-1}), the convective index (b), and the rainfall frequency (c, unit: %). The relative vorticity is averaged from 30-33°E, and the convective index and rainfall frequency are averaged from 28-33°E.

Figure 11

The TP low-pressure systems (vortex and shearline) occurring during 2 days before the start day to the end day of the cold (a) and warm (b) Yaan RREs. Colored lines show their pathways, and the triangles represent the start points.

Figure 12

The evolution of the area averaged ($30\text{-}33^\circ\text{N}$, $102\text{-}104^\circ\text{E}$) vorticity anomaly tendency (a, unit: 10^{-10} s^{-1}) and its components (b-j). The horizontal axis represents the hours from the start of cold Yaan RREs, and the y axis represents the height.