Impact of Dynamic and Thermal Forcing on the Intensity Evolution of the Vortices over the Tibetan Plateau in Boreal Summer

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Abstract: The Tibetan Plateau Vortex (TPV) is one of the main weather systems causing heavy rainfall over the Tibetan Plateau in boreal summer. Based on the second Modern-Era Retrospective Analysis for Research and Applications (MERRA-2) reanalysis datasets provided by the National Aeronautics and Space Administration (NASA), 8 cases of TPV over the Tibetan Plateau generated in June-August with a lifetime of 42 hours are composited and analyzed to reveal the impact of dynamic and thermal forcing on the intensity evolution of TPVs. The results are as follows. (1) The TPVs appear obviously at 500 hPa and the TPVs intensity (TPVI) shows an obvious diurnal variation with the strongest at 00LT and the weakest at 12LT (LT=UTC+6h). (2) A strong South Asia High at 200 hPa as well as a shrunken Western Pacific Subtropical High at 500 hPa provide favorable conditions for the TPVI increasing. (3) The vorticity budget reveals that the divergence is indicative of the variation of the TPVI. The TPVI decreases when the convergence center at 500 hPa and the divergence center at 200 hPa lie in the east of the TPVs center and increases when both centers coincide with the TPVs center. (4) Potential vorticity (PV) increases with the enhancement of the TPVI. The PV budget shows that the variation of the TPVI is closely related to the diabatic heating over the Tibetan Plateau. The increased sensible heating and radiative heating in the boundary layer intensify the ascent and latent heating release. When the diabatic heating center rises to 400 hPa, it facilitates the development of the TPVs.

Key words: Tibetan Plateau Vortex; intensity; dynamic composite method; diabatic heating

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1 INTRODUCTION

The Tibetan Plateau (TP) is located in central-eastern Asia, with an average altitude of more than 4,000 m. The TP can influence the formation of the climate in East Asia and even in the northern hemisphere through both dynamic and thermal forcing (Curio et al. [1]; Liu et al. [2]; Wu et al. [3]; Yao et al. [4-5]; Zheng et al. [6]). Low-pressure system is the main weather system producing rainfall over the TP, and under favorable circulation conditions, its development and eastward movement can trigger severe weather such as torrential rain downstream of the TP, and even in the middle-lower reaches of the Yangtze River, the Huang-Huai River Basin and other regions in eastern China (Fu et al. [7]; Jian et al. [8]; Wan et al. [9]; Yao et al. [10]; Zhang et al. [11]; Zhao et al. [12]). Exploring the activity rules of the low-pressure systems is of great significance for investigating the weather process in the TP and the areas to its east. The Tibetan Plateau vortex (TPV) is the main low-pressure system producing rainfall over the TP in boreal summer (Li et al. [13]). The TPV is a shallow meso-scale system in the boundary layer with typical scales of 400–500 km in the horizontal direction, and 2–3 km in the vertical direction (Li [14]; Zhang et al. [15]). It usually has a lifetime of 1–3 days (Li [14]; Zhang et al. [15]) and occurs more often during May to September with a peak in June (Ye and Gao [16]).

The impacts of large-scale circulation and the atmospheric dynamic forcing on the development and eastward movement of TPVs have been of great concern for a long time (Li et al. [17]). In boreal summer, the South Asia High covers the TP at 200 hPa (Liu et al. [18]; Wei et al. [19]), which strengthens when the TPVs occur, providing favorable dynamic condition of divergence at the upper layer (Li et al. [20]; Zhao [21]; Li et al. [22]). The strong divergence center on the north side of the westerly jet axis at 300 hPa can guide the TPVs to move eastward along the north side of the jet (Sun and Chen [23]). The TPVs would strengthen when the convergence center at 500 hPa lies below the entrance of the westerly jet at 200 hPa (Zhang et al. [15]; Li et al. [20]). In addition, as another important factor influencing the development of TPVs, the dynamic pumping, i.e., Ekman Pumping, in the boundary layer is closely related to the distribution of the horizontal divergence over the
TP. When the pumping is “up”, it is beneficial for the development of the TPVs (Li and Xu [24]).

The TP is the highest plateau with complicated topography in the world, and its thermal effects have a significant impact on the atmospheric circulation in East Asia (Yao et al. [4, 5]; Zheng et al. [6]; Zhang et al. [23]). The positive vorticity source caused by the heating over the TP is important for the generation and maintenance of the cyclonic circulation at the lower layer in boreal summer (Liu et al. [26]). Some researchers have pointed out that latent heating release plays an important role in the generation and development of the TPVs (Dell and Chen [27]; Wang [28]; Li et al. [29]). The sensitivity test shows that if there is no latent heating release, the TPVs intensity (TPVI) decreases significantly or the TPVs cannot even generate (Wang [29]). The high-value center of the atmospheric apparent heat source ($Q_1$) is located at 400 hPa, which is mainly derived from the latent heating release. Under the condition of static stable stratification, the vertical gradient of diabatic heating promotes the development of the TPVs that lie below the maximum of diabatic heating (Zheng et al. [30]). And the horizontal gradient of diabatic heating promotes the development of the TPVs that lie on the right side of the vertical shear of the horizontal wind (Zheng et al. [30]). Besides, attention should also be paid to the role of sensible heating over the TP in the development of TPVs (Wu et al. [3]; Zhao [31]; Shen et al. [32]; Li et al. [32]; Wen et al. [33]; Zhang and Li [34]; Zhang and Li [35]; Zhang et al. [36]). Without sensible heating, there will be few intersections between the small-scale and large-scale systems such as CISK (the conditional instability of second kind) (Wu et al. [7]). In previous studies, it has been found that the sensible heating flux in boreal summer is highly correlated with the frequency of TPVs generation in the same period (Zhang and Li [34]), that is, the increase of sensible heating is beneficial to the deepening of low-pressure systems and the generation of TPVs (Wen et al. [33]; Zhang et al. [36]). Moreover, especially when sensible heating is combined with appropriate weather conditions and terrain, it will play a great role in the development of TPVs (Shen et al. [31]). There are relatively few studies for the effect of radiative heating on TPVs, but it is certain that the radiative heating is also of great significance to the change of vorticity (Zhao [31]; Ge et al. [17]).

As a typical weather system in the boundary layer over the TP, the TPVs are significantly affected by the diabatic heating and the diurnal variation of the boundary layer (Zhang et al. [13]; Liu et al. [18]; Liu et al. [26]). The relative vorticity at 500 hPa over the TP is the weakest at 12LT (LT=UTC+6h) and the strongest at 00LT during a day (Chow and Chan [39]). Correspondingly, the occurrence frequency of the TPVs has a diurnal variation with its minimum from early morning to noon (06–12LT) and maximum from evening to midnight (18–00LT) (Zhang and Li [34]; Feng et al. [39]; Li et al. [40]). Li et al. further explored the diurnal variation of the intensity of nascent TPVs and found that the intensity of nascent TPVs generating at 18–00LT is relatively stronger with the most occurrence in the central and western TP, while that at 06–12LT is relatively weaker with the most occurrence in the central and eastern TP [29]. This is mainly due to the enhancement of convergence at 500 hPa and divergence at 200 hPa at night with the strengthening in the ascent as well as the increase of latent heating release (Li et al. [29]; Li et al. [41]).

It can be seen that as a weather system in the boundary layer, the development and evolution process of TPVs are inevitably affected by the diurnal variation of the dynamic and thermal conditions in the boundary layer. However, in past studies, there are few research focusing on the influence of the diurnal variation of the boundary layer on the TPVs (Li et al. [29]). In addition, although many studies have been carried out on the mechanisms for the generation and movement of TPVs, what the difference between the dynamic and thermal forcing in different cases or in different stages of the same case is and how they affect the TPVs respectively still remain elusive (Chen et al. [42]).

To improve the understanding of the generation and development of TPVs and provide better theoretical support for weather forecasts over the TP, 8 cases of TPV with the same lifetime are composited and analyzed to reveal the impact of dynamic and thermal forcing on the evolution of TPVI. The evolution process of TPVI is described in section 3. The dynamic forcing is analyzed by using vorticity budget and the evolution of the potential vorticity (PV) and PV budgets are studied to explore the effect of diabatic heating on TPVI in section 4, and a possible mechanism for the evolution of TPVI is proposed in this section. Conclusions and discussions are given in section 5.

2 DATA AND METHODS

2.1 Data

The data used in this paper is the second Modern-Era Retrospective Analysis for Research and Applications (MERRA-2), provided by the National Aeronautics and Space Administration (NASA)’s Global Modeling and Assimilation Office (Gelaro et al. [43]; Ma et al. [44], with a horizontal resolution of 0.625°×0.5°, for the period of June-August during 2004–2016. The data involves surface latent heating flux, surface net shortwave radiation flux and surface sensible heating flux from the dataset of “tavg1_2d_ind_Nx (M2T1NXLND): Land Surface Diagnostics” (DATA1); horizontal wind ($u$ and $v$), vertical wind in the pressure coordinate ($\omega$), air temperature ($T$) and height ($H$) from the dataset of “inst3_3d_asm_Np (M2I3NPASM): Assimilated Meteorological Fields” (DATA2). The data in DATA1 are only at surface level with an interval of 1 hour, and the data in DATA2 has 27 vertical levels with
an interval of 3 hours.

The cases selected in this study are all from the Yearbook of Tibetan Plateau Vortex and Shear Line (Yearbook) accomplished by the Institute of Plateau Meteorology of China Meteorological Administration \(^{[49]}\). 2.2 Dynamic Composite method

The selected cases used for the dynamic composite analysis need to meet the following four conditions.

1) The TPV which is the most obvious at 500 hPa should be recorded in both the Yearbook and MERRA-2.
2) The TPV causes precipitation over the TP.
3) The starting and ending time of the lifetime for all TPVs should be consistent with each other. The lifetime is defined as the period from the first occurrence of a closed low-value contour line at 500 hPa with both the high-value center of the vorticity and cyclonic wind shear, to the last occurrence of the closed contour line.
4) The TPV should have a lifetime of more than 24 hours and its center should be located on the main body of the TP (with an altitude above 3,000 m) during its entire lifetime.

According to the above criteria, 8 cases of TPV are finally selected (Table 1 and Fig. 1). All selected TPVs generate at 21LT (LT=UTC + 6h. From this point forward, data and time will be expressed as DN\NNLT, e.g., D1\21LT denotes 21LT on the first day of the TPV lifetime and D2\12LT denotes 12LT on the second day of the TPV lifetime) lasting for 42 hours and 15 times in total, and then dissipate at D3\15LT.

In this paper, considering the movement of TPVs, we choose to use a dynamic composite method (Li et al. \(^{[14]}\); Li et al. \(^{[17]}\); Li et al. \(^{[19]}\)). The TPV center at 500 hPa is taken as the origin of coordinate, which moves along with the TPV, and then the average value of each physical quantity at each point in the 8 cases is calculated, which has the same distance to the TPV center.

| Number | Case 1 | Case 2 | Case 3 | Case 4 | Case 5 | Case 6 | Case 7 | Case 8 |
|--------|--------|--------|--------|--------|--------|--------|--------|--------|
| Year   | 2004   | 2005   | 2005   | 2005   | 2012   | 2015   | 2015   | 2016   |
| Month  | 7      | 6      | 7      | 7      | 8      | 8      | 7      |        |
| Day    | 23-25  | 17-19  | 20-22  | 27-29  | 27-29  | 05-07  | 17-19  | 03-05  |
| Time   | D1\21LT – D3\15LT |

Figure 1. Trajectory of each TPV at 500 hPa. The shading is the altitude (units: m) and the gray thick solid line is the isoline at 3000 m; the polylines of different colors represent the moving tracks of different TPVs; the solid dot denotes the position of each TPV center at the initial time (D1\21LT) and the cross (+) represents the position of TPV centers every 3 hours.

2.3 Atmospheric diabatic heating

The atmospheric adiabatic heating consists of three parts: the sensible heating, the latent heating and the radiative heating. A reverse calculation method is used to calculate the atmospheric apparent heat source \(Q\) and the apparent moisture sink \(Q\) (Li et al. \(^{[17]}\); Li et al. \(^{[29]}\); Li et al. \(^{[40]}\); Yanai et al. \(^{[46]}\)), and the atmospheric diabatic heating rate \((\dot{Q})\) and latent heating rate \((\dot{Q}_{\text{lh}})\) is calculated through \(Q\) and \(Q\) as follows:

\[
\dot{Q} = C_s (\frac{\partial T}{\partial t} + \bar{v} \cdot \nabla T) + \omega (\frac{P}{P_0}) \frac{\partial \theta}{\partial p}
\]  

\[
\dot{Q}_{\text{lh}} = -L (\frac{\partial q}{\partial t} + \bar{v} \cdot \nabla q + \omega \frac{\partial q}{\partial p})
\]

\[
\dot{Q} = \frac{Q}{C_p}
\]
\[ Q_{DL} = \frac{Q_2}{C_p} \]  
\[ \langle Q_1 \rangle = \frac{1}{9} \int_{\phi_1}^{\phi_2} Q_1 d\phi \]  
\[ \langle Q_2 \rangle = \frac{1}{9} \int_{\phi_1}^{\phi_2} Q_2 d\phi \]

where \( Q \) is the diabatic heating rate, indicating the degree of diabatic heating per unit time, \( Q_{DL} \) represents the degree of latent heating release from the condensation of water vapor per unit time. \( C_p \) = 1004.8416 J·kg\(^{-1}\)·K\(^{-1}\). \( \kappa \) is the specific heat at constant pressure. \( T \) is the temperature. \( \vec{v} \) is the horizontal wind vector. \( \theta \) is the potential temperature. \( L \) is the coefficient of latent heating and \( q \) is the specific humidity. \( \omega \) is the vertical velocity in the pressure coordinate (\( P \) coordinate). In Eq. (5) and Eq. (6), \( P \) denotes the pressure at the top of atmosphere (taking the value of 100 hPa), \( P_o = 1000 \) hPa and \( P_r \) is the surface pressure.

3 EVOLUTION OF TPVI

In this paper, the vertical component of the relative vorticity (\( \zeta \)) at the composite center of TPVs is taken to describe the TPVI:

\[ \zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \]

where \( u \) and \( v \) represent the zonal and meridional wind speed, respectively.

In Fig.2a, during the lifetime, \( \zeta \) near the composite center of TPVs reaches the maximum at 500 hPa and the positive region can stretch to 350 hPa with a thickness of 3,500 m. Below the level of 450 hPa, the TPVI shows an obvious diurnal variation, which experiences four stages of the first increase (D1 \& 21LT-D2 \& 06LT), the first decrease (D2\& 06LT-D2\& 12LT), the second increase (D2\& 12LT-D3 \& 00LT), the second decrease (D3 \& 00LT-D3 \& 15LT). At the initial time (D1 \& 21LT), the range of \( \zeta \) greater than 2.0 \times 10^{-5} \text{s}^{-1} (\( \zeta = 2.0 \) in short) stretches to about 470 hPa. Subsequently, the TPVI increases and the range of \( \zeta \) stretches to 450 hPa. After D2\& 06LT, the TPVI begins to decrease. The stretching height of \( \zeta \) gradually reduces to 480 hPa at D2\& 12LT, which is the weakest moment of TPVI in lifetime. After that, the TPVI begins to increase again during D2\& 18LT-D3 \& 00LT, during which the range of \( \zeta \) can once stretch up to 400 hPa. In addition, from D2\& 21LT, the region of \( \zeta \) greater than 4.0 \times 10^{-5} \text{s}^{-1} appears at 500-450 hPa, and stretches up to 450 hPa at D3\& 00LT, indicating that the TPVI enters the strongest moment in lifetime. After D3\& 00LT, the stretching height of \( \zeta \) greater than 4.0 \times 10^{-5} \text{s}^{-1} reduces sharply and disappears at D3 \& 06LT. The stretching height of \( \zeta \) will stay at about 400 hPa until the TPVs dissipate at D3\& 15LT.

It is found that (Fig. 2b) the evolution of \( \zeta \) at 500 hPa shows a bimodal distribution: during D1\& 21LT – D2\& 06LT, \( \zeta \) increases and reaches the first peak of 3.5 \times 10^{-5} \text{s}^{-1} at D2\& 06LT; while during D2\& 06LT – D2\& 12LT, \( \zeta \) suddenly decreases, reaching the minimum of 2.0 \times 10^{-5} \text{s}^{-1} at D2\& 12LT in lifetime; after that, \( \zeta \) increases again and reaches the second peak and the maximum of 5.5 \times 10^{-5} \text{s}^{-1} at D3\& 00LT in lifetime. After D3\& 00LT, \( \zeta \) decreases gradually. Finally, the TPVs dissipates when the cyclonic circulation and closed contour line disappear after D3\& 15LT.

From the above analysis, it can be seen that the TPVI shows an obvious diurnal variation during the evolution of TPVs. The TPVI decreases during 09-12LT, while it increases significantly during 21-00LT. The dynamic and thermal effects leading to the evolution of TPVI will be explored through diagnostic analysis in the next section.

\[ \text{AVE}=3.4 \times 10^{-5} \text{s}^{-1} \]

Figure 2. (a) Height-time cross sections of the area-averaged (in the range of 9 × 9 grid point near the composite center of TPVs) relative vorticity (\( \zeta \), units: 10^{-5} \text{s}^{-1}), and (b) time series of the area-averaged relative vorticity departure at 500 hPa.

4 DIAGNOSTIC ANALYSIS FOR THE EVOLUTION OF TPVI

4.1 Circulation

Comparing the atmospheric circulations at the strongest time (D3\& 00LT) and weakest time (D2\& 12LT) of the TPVI (Fig. 3), it can be seen that the TPVs at 500 hPa develop from the cut-off lows, and the composite center of the TPVs at 200 hPa is on the south of the westerly jet and north of the South Asia high. At D3\& 00LT (Fig. 3b), the contour lines near the composite center of the TPVs at 500 hPa are dense with the closed
center of 5,825 m. There is a cyclonic circulation near the composite center of the TPVs with the scope of 40 grid points (25°) from east to west and 20 grid points (10°) from north to south. Straight westerlies lie to the north of the composite center of the TPVs, while the area to the southeast is controlled by the Western Pacific subtropical high. The southerly wind in the peripheral region of the Western Pacific subtropical high gradually turns into westerly wind at 20 grid spacings to the east of TPVs. At 200 hPa, the composite center of TPVs is located in a region with extensive anti-cyclonic wind shear on the north of South Asia High (Fig. 3d). Compared with the strongest moment of TPVI, at D2/12LT (Fig. 3a), the contour lines near the center at 500 hPa is relatively sparse with the closed center of 5,830 m. The scope of the cyclonic circulation significantly reduces while the Western Pacific subtropical high to the southeast side intensifies and the range expands. The height at 200 hPa of the two moments is similar (Fig. 3c-d). It should be noted that the horizontal wind velocity near the composite center of TPVs at the upper and lower layers at D3\00LT are stronger than that at D2\12LT. Especially at 500 hPa, the wind speed difference in the southeast of the composite center of TPVs exceeds 6 m s\(^{-1}\), which is the inevitable result of TPVs deepening and contour densifying.

From the above analysis, at the strongest moment of TPVI, the South Asia High at 200 hPa is strong, whereas the Western Pacific subtropical high to the southeast of TPVs at 500 hPa is relatively weak with a shrunken scope. This circulation pattern provides favorable conditions for the development of cyclonic vorticity and the deepening of TPVs.

Figure 3. (a) Composites of heights (contour, units: m). The contour interval at 500 hPa is 10 m below 5820 m and 5 m above 5820 m, and 50 m at 200 hPa, respectively) and winds (vector, shading is the horizontal wind difference between D3\00LT and D2\12LT, namely \(\vec{v}_t(D3\00LT)-\vec{v}_t(D2\12LT)\), units: m s\(^{-1}\)) at 500 hPa at (a) D2\12LT and (b) D3\00LT, and at 200hPa at (c) D2\12LT and (d) D3\00LT. (0,0) is the composite center of TPVs, and the x-axis and y-axis represent the number of grid points from the center.

4.2 Effect of dynamic forcing on the evolution of TPVI

In this section, the classic vorticity budget equation is used to explore the effect of atmospheric dynamic forcing on the evolution of TPVI. The simplified vorticity budget equation after the magnitude analysis is as follows:

\[
\frac{\partial \zeta}{\partial t} = -(\vec{v} \cdot \nabla)\zeta - (\zeta + f) \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right)
\]  

(8)

The left end of Eq. (8) is the tendency term, and the right end is the advection term and the divergence term.

As can be seen from Fig. 4, at the lower layer, the tendency term shows an obvious diurnal variation, i.e., positive at 15-21LT and negative at 03-12LT. It reaches its maximum 6 hours earlier than the vorticity. At the initial stage of TPVs (D1\12LT-D2\00LT), the tendency term is positive, indicating that the TPVs will develop and strengthen. At D2\03LT-D2\12LT, it turns to a negative value. During this period, the development of TPVs gradually slows down, and then weakens. At D2\
12LT, it reaches the minimum TPVI (Fig. 2). After that, the tendency term changes to positive during D2\15LT- D2\21LT, and gets the maximum at D2\18LT, promoting the increase of the TPVI. After D3\00LT, the tendency term is always negative until the TPVs dissipate. The occurrence time of the positive and negative center of the tendency term in the upper layer is about 9 hours later than that in the lower layer.

Then pay attention to the advection term. Below 450 hPa, there is a big difference between the distribution of the advection term and the tendency term. The advection term is negative except during D2\03LT-D2\06LT, and the absolute value is the largest at D2\21LT, indicating that the horizontal vorticity transport is unfavorable for the development of TPVs. Above 400 hPa, the advection term is mainly positive, and its distribution corresponds to the tendency term.

As can be seen from Fig. 4b, the divergence term is positive below 450 hPa and always contributes positively to the development of TPVs. Its value has obvious daily variation, which is smaller during 03-12LT with the minimum of $5 \times 10^{-16}$s$^{-2}$ and larger during 15-00LT with the maximum of $40 \times 10^{-16}$s$^{-2}$. The distribution of the divergence term is related to the divergence. The convergence maintains at 500-400 hPa, and the intensity of convergence shows an obvious diurnal variation too, that is, the convergence intensifies during 18-21LT and weakens at 06LT. The occurrence of convergence centers at 350-300 hPa is 6 hours later than that at 500 hPa. The atmosphere above 250 hPa is always characterized by divergence with the center at around 150 hPa. The intensity of the divergence at 150 hPa decreases during D2\06LT-D2\15LT.

According to the above analysis, below 400 hPa, the divergence term decreases, and the convergence weakens during 03-12LT, which reduces the conversion from convergence to vorticity, so the promotion to the development of TPVs is limited. At the same time, the transport of low vorticity to the center of TPVs decreases and the absolute value of negative advection term decreases, which weakens the negative effect of advection on TPVs. The tendency term is negative now. Above 300 hPa, the divergence term is negative with a small absolute value, and the center of TPVs is controlled by divergence, which transports the negative vorticity outwards. Therefore, divergence is the most important dynamic forcing that affects the TPVI.

**Figure 4.** Height-time cross sections of the area-averaged (in the range of 9 x 9 grid point near the composite center of TPVs). (a) The tendency term (shading, units: $10^{-16}$s$^{-2}$) and the advection term (contour, units: $10^{-16}$s$^{-2}$) and (b) the divergence term (shading, units: $10^{-16}$s$^{-2}$) and divergence (contour, units: $10^{-6}$s$^{-1}$).

It is worth noting that the occurrence of the maximum (minimum) $\zeta$ at 500 hPa is 6 hours later than that of the maximum (minimum) convergence, indicating that the divergence is indicative of the change in the TPVI. Comparing the distributions of the divergence and the vertical velocity at the upper and lower layer at D2\06LT (6 hours before the weakest time of the TPVI and the weakest time of convergence at 500 hPa), D2\12LT (the weakest time of TPVI), D2\18LT (6 hours before the strongest time of TPVI and the strongest time of convergence at 500 hPa) and D3\00LT (the strongest time of TPVI) (Fig. 5), it can be seen that at the four moments near the composite center of TPVs, there is a convergence zone at 500 hPa and a divergence zone at 200 hPa, but the intensity and distribution of the convergence and divergence zones are quite different at each time. The ascent dominates at 500 hPa, which is related to the strong sensible heating effect over the TP in boreal summer. In the decrease stage at D2\06LT and D2\12LT for the TPVI, the convergence is weak near the composite center of TPVs at 500 hPa, and the convergence center is located in the east of TPVs (Fig. 5a-b). At 200 hPa, the TPVs locate in the weak divergence region between two strong divergence centers, with the intensity less than $0.4 \times 10^{-4}$s$^{-1}$ (Fig. 5e-f). The maximum ascent is located in the east of TPVs. It is worth noting that although the TPVI is the weakest at D2\12LT, it begins to increase at the next time. The convergence center at 500 hPa and divergence center at 200 hPa locate in the east of TPVs at D2\12LT, but they are closer to the composite center of TPVs than those at D2\06LT. In the increase stage at D2\18LT and D3\00LT for the TPVI, the convergence area near the composite center of TPVs at 500 hPa exhibits a zonal
distribution with the high-value coinciding with the composite center of TPVs, and the intensity could reach up to $-2.0 \times 10^{-3}$ s$^{-1}$ (Fig. 5c-d). At 200 hPa, the strong divergence center with the intensity of more than 1.2 $\times 10^{-5}$ s$^{-1}$ coincides with the composite center of TPVs (Fig. 5g-h). The maximum ascent is located near the composite center of TPVs. Although the TPVI is the strongest at $D3 \backslash 00$LT, it begins to decrease at the next time. The convergence center at 500 hPa and the divergence center at 200 hPa coincide with the composite center of TPVs at $D3 \backslash 00$LT, but they start to deviate eastward from the composite center of TPVs compared with those at $D2 \backslash 18$LT. By comparing the divergence distribution at the lower and upper layers at different stages in the lifetime of TPVs, it is found that the convergence centers and divergence centers are all located in the east of the composite center of TPVs at the decrease stage of TPVI, while all coincide with the composite center of TPVs at the increase stage of TPVI (figures not shown). Li et al. focused on the TPVs moving out of the TP and revealed that when the convergence center at 500 hPa and the divergence center at 200 hPa locate to the east of the TPVs center, it was conducive to the development and eastern movement of the TPVs$^{[22]}$. They also point out that compared with the eastward moving TPVs, the large value region of the convergence center at 500 hPa is closer to the center of the TPVs when the TPVs is on the main body of the TP. It is consistent with the conclusion obtained above.

![Figure 5](image)

Figure 5. Composites of divergence (shading, units: $10^{-3}$ s$^{-1}$) and vertical velocity (contour, units: $10^{-3}$ Pa s$^{-1}$) at 500 hPa at (a) $D2 \backslash 06$LT, (b) $D2 \backslash 12$LT, (c) $D2 \backslash 18$LT, (d) $D3 \backslash 00$LT, and at 200 hPa at (e) $D2 \backslash 06$LT, (f) $D2 \backslash 12$LT, (g) $D2 \backslash 18$LT, (h) $D3 \backslash 00$LT. (0,0) is the composite center of TPVs, and the x-axis and y-axis represent the number of grid points from the center.
The vorticity budget results show that the divergence term is an important factor affecting the local variation of vorticity, and the convergence intensity at 500 hPa is indicative to the variation of TPVI. The maximum (minimum) \( \zeta \) appears 6 hours later than the maximum (minimum) convergence. The TPVI decreases when the convergence at 500 hPa and divergence at 200 hPa center is in the east of the composite center of TPVs and increases when both two centers coincide with the composite center of TPVs.

4.3 Effect of diabatic heating on the evolution of TPVI

Since the diabatic heating plays an important role in the development and evolution of weather systems, it is necessary to investigate the changes of diabatic heating during the evolution of TPVI. In general, the value of total diabatic heating \( Q \) is the same as that of condensation latent heating \( Q_{\text{LH}} \), the distribution is similar, and the center of large value coincides; but \( <Q_1> \) and \( <Q_2> \) have different trend over time. The diabatic heating always heats the atmosphere except at D2/03LT. Although the whole process is accompanied by the precipitation and latent heating release, the latent heating is not as strong as the sensible heating and radiative heating at the lower layer during 06-12LT, so the maximum \( Q \) is located at the lower layer (Fig. 7). When the ascent intensifies at 18-00LT (figures not shown), latent heating release increases, and maximum \( Q \) rises to the middle layer. The maximum \( Q_{\text{LH}} \) is located at 450–350 hPa and exceeds 2.5K/3h at D2/21LT. The maximum \( Q_{\text{LH}} \) also shows a maximum of 2.5K/3h, and the range is greater than that of \( Q \). Due to the large ascent during D2/21LT – D3/06LT, the combined effects of cloud cover and evaporation cooling cause the surface radiative heating flux and sensible heating flux at D3/06LT are less than those at D2/06LT (Fig. 7). At this time, the release of latent heating of condensation increases compared with the same period of the previous day. Under the action of three diabatic heating method, the maximum of \( <Q_1> \) appears at D3/09LT. During 09-12LT, \( <Q_1> \) is larger than \( <Q_2> \), which is related to the effect of surface sensible heating during this period (Fig. 7). During 18-03LT, \( <Q_1> \) is larger than \( <Q_2> \), because the surface sensible heating and radiation heating is negative.

The potential vorticity (PV) is a physical quantity that contains the dynamic and thermodynamic characteristics of the atmosphere, and it is greatly affected by the diabatic heating at the lower layer. Previous studies have shown that PV can reflect the development and movement of TPVs well (Li et al. [17]; Li et al. [22], Li et al. [40]). In this paper, the influence of diabatic heating on the TPVI is discussed by studying the variation of PV.

The commonly used expressions of PV and its components in P coordinate system are as follows:

\[
PV = -g\left(f + \tilde{\zeta}\right)\frac{\partial \theta}{\partial p} + \frac{\partial v}{\partial p}\frac{\partial \theta}{\partial x} - \frac{\partial u}{\partial p}\frac{\partial \theta}{\partial y}
\]  \(\text{(9)}\)

\[
PV_x = \frac{\partial v}{\partial p}\frac{\partial \theta}{\partial x} - \frac{\partial u}{\partial p}\frac{\partial \theta}{\partial y}
\]  \(\text{(10)}\)

\[
PV_y = -g\left(f + \tilde{\zeta}\right)\frac{\partial \theta}{\partial p}
\]  \(\text{(11)}\)

\[
PV_v = -g\left(f + \tilde{\zeta}\right)\frac{\partial \theta}{\partial p}
\]  \(\text{(12)}\)

where \( f = 2\Omega \sin \phi \). \( PV_x \), \( PV_y \), and \( PV_v \) represent the \( X \) component, \( Y \) component (collectively referred to as PV baroclinic term) and \( P \) component (referred to as PV barotropic term) of PV, respectively. The diagnostic equation of PV budget in \( P \) coordinate system is as follows:

\[
\frac{\partial PV}{\partial t} = -\nabla \cdot \nabla PV + g\left(\nabla^2 \tilde{\zeta} + \tilde{F} \times \nabla \theta \right)
\]  \(\text{(13)}\)
The left end of Eq. (13) is the PV tendency term, and the right end is the PV advection term and the PV forced term, respectively. The forced term can be divided into two terms of the diabatic heating effect term and the friction effect term. $F$ is friction; $\xi_{a}$ is an approximation of three-dimensional absolute vorticity in $P$ coordinate. In this paper, only the PV advection term and the diabatic heating effect term on the PV tendency term are discussed.

Furthermore, the diabatic heating effect term can be divided into three parts:

$$
g \cdot \nabla (Q \xi_{a}) = g \frac{\partial u}{\partial p} \frac{\partial Q}{\partial x} - g \frac{\partial v}{\partial p} \frac{\partial Q}{\partial y} - g(f + \xi_{a}) \frac{\partial Q}{\partial p} \tag{14}
$$

From left to right, the right end of Eq. (14) is the $X$ component of diabatic heating effect term, the $Y$ component of diabatic heating effect term and the $P$ component of diabatic heating effect term.

As shown in Fig. 8a, the changes in PV at 500–400 hPa is consistent with that of $\xi$. The larger the PV value is, the larger the $\xi$ value will be (Fig. 2a). PV is always positive near the composite center of TPVs at 500 hPa (Fig. 8b). The PV baroclinic term is always negative with a small magnitude. The value of the PV baroclinic term decreases with the increase of the PV value, and the absolute value of PV at $\xi$ is larger than that of PV, which may be related to the larger zonal gradient of potential temperature. The PV barotropic term is equivalent to the magnitude of PV, and the variation trend is consistent with that of PV. Therefore, the change in PV at 500 hPa is mainly caused by the barotropic process which is consistent with the divergence diagnosis above.

![Figure 8](image1.png)

Figure 8. (a) Height-time cross sections of the area-averaged (in the range of 9 x 9 grid point near the composite center of TPVs) PV, and (b) time series of the area-averaged PV$_{x}$, PV, and PV$_{y}$ at 500 hPa. Units: for PV and PV$_{y}$ are PVU, where 1PVU $= 10^{-4}$ m$^2$ s$^{-1}$ K kg$^{-1}$; for PV$_{x}$ is $10^{-5}$ PVU, and for PV$_{y}$ is $10^{-6}$ PVU.

The diagnosis of the PV evolution at each time is conducted by using the Eq. (13), and the results are shown in Fig. 9. The change in diabatic heating effect term is the main reason of the change in PV tendency at 500–400 hPa. The PV advection term begins to play a leading role at about 400 hPa, especially at D2/21 LT, the large value of PV at lower layer are transported upward with a strong vertical velocity, and there is a center of the PV advection term at 350 hPa. The diabatic heating effect term and the PV advection term act together at the upper layers above 300 hPa.

![Figure 9](image2.png)

Figure 9. Height-time cross sections of the area-averaged (in the range of 9 x 9 grid point near the composite center of TPVs) PV tendency (PVT, shading), PV advection (PVA, black thick solid). The short dash denotes the negative value where the solid denotes the positive value (the same below), and the diabatic heating effects (DHE, red thin solid.). Units: PVU 3h$^{-1}$.
The PV diagnosis at 500 hPa is further analyzed in detail (Fig. 10). During the whole lifetime, the PV tendency term and the diabatic heating effect term show the same trend, but the absolute value of the diabatic heating effect term is larger with more dramatic change. The PV advection term always contributes negatively to the PV tendency term, and the absolute value is small with a gentle change. Among the three components of diabatic heating effect term, the variation trend of the $X$ component of diabatic heating effect term is opposite to that of the total diabatic heating effect term and the magnitude is small. The $Y$ component of diabatic heating effect term is also small in magnitude and negative except at D1 \ 21LT and D3 \ 15LT. The trend and magnitude of the $P$ component of diabatic heating effect term are similar to that of the total diabatic heating effect term, which plays a decisive role (Fig. 10b). The results show that the vertical gradient of $Q$ plays an important role in the variation of PV at 500 hPa. The PV tendency term and the diabatic heating effect term at 500 hPa show an obvious diurnal variation under the influence of diabatic heating in the boundary layer. (Fig. 10a). They are positive during D1\21LT-D2\03LT and D2\15LT-D3\03LT, indicating that the dynamic and thermal conditions are favorable for the increase of PV and the development of TPVs, that is, due to the strong ascent and the latent heating release, the maximum $Q$ uplifts to 400 hPa with $\frac{\partial Q}{\partial p} < 0$ at 500 hPa; while they are negative during D2\06LT – D2\12LT and D3\06LT – D3\12LT, indicating that the dynamic and thermal conditions are unfavorable for the increase of PV and the development of TPVs, that is, the surface radiative heating flux and sensible heating flux increase significantly (Fig. 7), and the maximum $Q$ is located at the lower layer with $\frac{\partial Q}{\partial p} > 0$ at 500 hPa. Both the PV tendency term and the diabatic heating effect term reach the minimum of the day at 09LT. The absolute values of the PV tendency term and the diabatic heating effect term at D2\09LT are smaller than those at D3\06LT, because the TPVs have developed during D2\18LT – D3\00LT, and the latent heating lifts the diabatic heating center up to the middle layer, whereas the vertical gradient of $Q$ is relatively small at D3\09LT. After D3\12LT, the PV tendency term changes to positive but the value is small. The TPVs eventually dissipate under the combined effects of the friction and the negative contribution of the PV advection term.

![Figure 10](image-url) (a) Time series of the area-averaged (in the range of 9 × 9 grid point near the composite center of TPVs) PV tendency (PVT, solid), PV advection (PVA, short dashed), and diabatic heating effects (DHE, dotted) at 500 hPa, and (b) time series of the area-averaged $X$ component of diabatic heating effects (DHEX, solid), $Y$ component of diabatic heating effects (DHEY, dotted) and $P$ component of diabatic heating effects (DHEP, short dashed) at 500 hPa. Units: for PVT, PVA, DHE and DHEP are PVU 3h$^{-1}$; for DHEX, DHEY are 10$^3$ PVU 3h$^{-1}$.

The variation of PV at the lower layer is very sensitive to the diabatic heating. The strong cooling near the surface over the TP at night (D1\21LT – D2\03LT) leads to the increase in the static stability and PV (Fig. 7 and Fig. 8a). During D2\06LT – D2\12LT, the surface radiative heating flux and sensible heating flux increase gradually while the static stability decreases rapidly which leads to the decrease in PV and $\zeta$. Similar to the situation during D1\21LT-D2\03LT, the cooling effects at night result in the increase of the static stability and PV at D2\15LT-D3\03LT. Moreover, the increased PV leads to the intensified of the ascent and latent heating release, and then the maximum $Q$ lifts up to 400 hPa; the evaporation of precipitation further leads to the decrease in $Q$ at the lower layer. The vertical gradient of $Q$ increases, and the static stability at 500–400 hPa continues to increase which leads to the strongest PV and TPV1 at D3\00LT-D3\03LT. After D3\12LT03, the variation of diabatic heating near the surface repeats the cycle of the previous day, therefore, PV and TPVI decrease.

To sum up, the diabatic heating is of great significance to the change in TPVI. The non-uniform distribution of $Q$ in the vertical direction is closely related to the diurnal variation of the PV and TPVI at 500 hPa. When the maximum $Q$ is located at 400 hPa, it
is favorable for the increase of TPVI.

4.4 Mechanisms for the evolution of TPVI

According to the above diagnostic analysis, a possible mechanism for the evolution of TPVI is presented in this section (Fig. 11). During D2\00LT – D2\06LT, the cooling near the surface layer over the TP is strong at night, so the atmosphere near the surface is rather colder and the isentropic surface becomes upper-convex (Fig. 11b). This situation would lead to an anomalous low-level divergence disturbance (a weakened convergence in the original field, Fig. 4) as well as an anomalous descent (a weakened ascent in the original field). Meanwhile, an anomalous convergence disturbance is generated at the upper layer of 150 hPa. During this period, the dynamic and thermal conditions are unfavorable for the development of TPVs, thus the TPVI turns to decrease from D2\06LT (Fig. 2b). During D2\06LT – D2\12LT, the surface radiative heating flux and sensible heating flux increase significantly (Fig. 7), so the atmosphere near the surface warms up and the isentropic surface turns to be lower-convex, breaking the decreasing mechanism of TPVI (Fig. 11b). During D2\15LT – D2\18LT, the warm atmosphere near the surface results in an obvious lower-convex isentropic surface. This situation forces an anomalous low-level convergence disturbance (an enhanced convergence in the original field, Fig. 4) and an anomalous ascent (an enhanced ascent in the original field), and so the TPVI turns to increase. The anomalous ascent could stretch to 300 hPa at D2\21LT. Massive latent heating is released and the maximum $Q$ greater than 2.5K/3h rises up to 430 – 350 hPa (Fig. 6), which warms up the atmosphere at the middle layer. Therefore, the pressure would increase at the upper layer, thus generating an anomalous divergence disturbance center at 150 hPa. The enhanced divergence at the upper layer further reduces the pressure and intensifies the convergence at the lower layer. The TPVs intensify sharply and TPVI increases to the strongest at D3\00LT (Fig. 2b). During D3\03LT-D3\09LT, the cooling in the lower levels over the TP forces an anomalous low-level convergence disturbance. Because of the anomalous convergence center caused by the heating at the middle layer, there is a weak anomalous descent below 350 hPa and an anomalous ascent above 350 hPa. The development mechanism of TPVs is destroyed, thus the TPVI decreases and finally the TPVs dissipate.

Overall, the diabatic heating plays an important role in the evolution of TPVI. The increased sensible heating and radiative heating in the boundary layer intensify the ascent. Latent heating release warms up the atmosphere at the middle layer causing the increased pressure and enhanced divergence at the upper layer, which further reduces the pressure and intensifies the convergence at the lower layer. As a result, the TPVs develop and the TPVI increases.

Figure 11. (a) Height-time cross sections of the area-averaged (in the range of $9 \times 9$ grid point near the composite center of TPVs) divergence departure (contour, units: $10^{-6}$ s$^{-1}$) and vertical velocity departure (shading, units: $10^{-2}$ Pa s$^{-1}$), and (b) height-time cross sections of the area-averaged potential temperature (contour and shading, units: K).
5 CONCLUSIONS AND DISCUSSION

In this paper, based on MERRA-2 reanalysis dataset, 8 cases of TPV are selected, which are generated at D1\21LT, lasting for 42 hours, then dying at D3\15LT, and have not moved out of the TP during the lifetime, to reveal the impact of dynamic and thermal forcing on the intensity evolution for TPVs through composited diagnosing. The main conclusions are as follows.

(1) In the 42-hour lifetime, the positive vorticity in the TPVs center can extend to 350 hPa in the vertical direction, but the maximum is at 500 hPa. The TPVI shows an obvious diurnal variation with the strongest at 00LT and the weakest at 12LT.

(2) At the strongest moment of TPVI, the South Asia High at 200 hPa is strong, and the Western Pacific Subtropical High on the southeast of the TPVs at 500 hPa is weak, providing favorable conditions for TPVI increasing.

(3) The diagnosis of the vorticity budget reveals that the divergence term is the most important factor affecting the local variation of vorticity. Although the convergence at 500 hPa will make the peripheral low-value vorticity flow toward the TPVs center, the convergence of airflow leads to the increase of the local vorticity under the Coriolis effect. The divergence is indicative to the variation of TPVI, and the maximum (minimum) vorticity at 500 hPa appears 6 hours later than the maximum (minimum) convergence. The TPVI decreases when the convergence at 500 hPa and divergence at 200 hPa center in the east of TPVs center and increases when both the two centers coincide with TPVs center.

(4) During the evolution of TPVs, PV increases with the development of TPVs. Results from the PV diagnosing show that the diabatic heating over the TP plays an important role in the evolution of TPVI. The increased sensible heating and radiative heating in the boundary layer intensify the ascent. Latent heating release warms up the atmosphere at the middle layer causing the increased pressure and enhanced divergence at the upper layer, which would further reduce the pressure and intensify the convergence at the lower layer. As a result, the TPVs develop with the increase of TPVI.

Compared previous previous studies, this paper explores the evolution mechanism of TPVI by using the dynamic composite method. Therefore, the conclusions are universal to some extent. In addition, this paper focuses on the effect of diurnal variation of the boundary layer on the development of TPVs. The difference of dynamic and thermal effect, the interaction between them and how they affect the TPVI are investigated in different stages of TPVs by using the vorticity budget equation and potential vorticity budget equation. A possible mechanism of TPVI evolution is given.

However, the study in this paper is conducted on specific cases of TPVs which generate at night in local time and dissipate at noon. Whether the evolution of TPVI with different lifetime follows the same rules remains to be further investigated.

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