Mechanism of Atmospheric Diabatic Heating Effect on the Intensity of Zonal Shear Line Over the Tibetan Plateau in Boreal Summer

Xiuping Yao1,2, Qiaohua Liu1,2,3, Shuo Zhang4, and Jiali Ma1,2,5

1China Meteorological Administration Training Centre, Beijing, China, 2State Key Laboratory of Severe Weather, Chinese Academy of Meteorological Sciences, Beijing, China, 3School of Atmospheric Science, Nanjing University of Information Science & Technology, Nanjing, China, 4Beijing Institute of Applied Meteorology, Beijing, China, 5University of Chinese Academy of Sciences, Beijing, China

Abstract The zonal shear line (ZSL) over the Tibetan Plateau (TP) is one of the most crucial synoptic systems inducing precipitation over the TP in boreal summer. However, few studies have comprehensively explored the thermal mechanism of the intensity evolution of the ZSL. In this study, the mechanism of the atmospheric diabatic heating (hereafter referred to as diabatic heating) effect on the intensity of the ZSL was explored by the methods of composite and diagnostic analysis. Based on the fifth generation European Centre for Medium-Range Weather Forecasts atmospheric reanalysis of the global climate hourly data sets from June to August during 1980–2019, 11 cases of the ZSL were selected by using the objective identification method. Results suggest that a close relationship exists between the ZSL's intensity and the 10-h-earlier vertically integrated diabatic heating, with a high correlation coefficient of 0.81. The intensity of the vertically non-uniform diabatic heating near the ZSL enhances (weakens), it will be favorable (unfavorable) to enhance the ZSL's intensity.

1. Introduction

The Tibetan Plateau (TP) is “the Roof of the World,” “the Water Tower of Asia,” and is known as the “Third Pole” (Qiu, 2008). With an average altitude of more than 4,000 m, the TP covers one-quarter of China’s land area and has an essential impact on the atmospheric circulation patterns and the weather and climate of China, East Asia, and even the world (Abe et al., 2013; Wu & Liu, 2016; Xu et al., 2008; Ye, 1988; Zhou et al., 2009). The unique cyclonic low-pressure synoptic systems, including the shear line over the TP (TPSL) and the plateau vortices (The Tibetan Plateau Science Research Group, 1981), are usually formed due to TP’s special geography and topographic environment. They often lead to severe weather events such as heavy rainfalls in both the TP and the downstream regions, causing heavy socio-economic losses (Li & Li, 2017a, 2017b). Many studies have shown that the TPSL is more important because the TPSL tends to induce the plateau vortices and provides an essential background for plateau vortex initiation (Du et al., 2020; Li et al., 2011; Yao et al., 2014).

Research on the TPSL began in the 1950s (Ye et al., 1957). The TPSL usually refers to the convergence line of the wind direction observed by at least three meteorological stations at 500 hPa (The Tibetan Plateau Science Research Group, 1981). Based on the trend pattern, the TPSL can be divided into two kinds, the zonal shear line (ZSL) and the meridional shear line over the TP (The Tibetan Plateau Science Research Group, 1981; Yao, Zhang, & Ma, 2020). The ZSL is one of the most crucial synoptic system over the TP in boreal summer (He & Shi, 2011). The high-frequency axis of the ZSL is located near 33°N. Moreover, the ZSL has a significant effect on precipitation over the TP as the correlation between precipitation and the ZSL in boreal summer can exceed 0.6 (X. Zhang et al., 2016).

As we know, the atmospheric diabatic heating (hereafter referred to as diabatic heating) consisting of radiative heating, sensible heating, and latent heating, which makes the TP behave as a powerful heat source.
in boreal summer (Yao, Yan, & Zhang, 2019; Ye et al., 1957). The latent heating of condensation during precipitation can reach the same magnitude as sensible heating in boreal summer, which makes a remarkable enhancement of TP diabatic heating and creates a significant temperature gradient between the TP and the surrounding regions (Ding, 1992; Flohn, 1957; Ye et al., 1957). As a result, the atmospheric circulation and climate in boreal summer are significantly affected (Duan et al., 2020; Duan & Wu, 2005; Y. M. Liu et al., 2012; G. Liu et al., 2009; B. Wang et al., 2008; Wu & Liu, 2016; Zhao et al., 2007). Besides, the diabatic heating of the TP also plays a crucial role in the formation and development of weather systems over the TP (Chen & Li, 2014; Liu & Wu, 2000; Wu et al., 2018). Studies have shown that diabatic heating, a conducive factor to convective instability and convective activities (Guan et al., 2018; Yanai & Li, 1994), is also one primary condition for the formation of the ZSL in boreal summer (Yao, Zhang, & Yan, 2019). The intensity of diabatic heating also influences the intensity of the ZSL and therefore plays an essential role in the intensity evolution of the ZSL (The Tibetan Plateau Science Research Group, 1981; Yao et al., 2014).

According to previous studies, a ZSL may vary considerably in the structure during the evolution process (Luo & Li, 2018; S. Zhang et al., 2019). So, what is the relationship between the intensity of ZSL and diabatic heating? What is the mechanism of the atmospheric diabatic heating effect on the intensity evolution of the ZSL? These two questions have not been answered to date. In addition, previous studies on the structural characteristics of the ZSL were mainly based on the results of a single experiment or case. Therefore, the understanding of the ZSL is inevitably limited, and the research of the diabatic heating effect on the ZSL is not comprehensive.

Based on the fifth generation European Centre for Medium-Range Weather Forecasts (ECMWF) atmospheric reanalysis of the global climate (ERA5) hourly data sets, the study explores the relationship between the ZSL’s intensity and the intensity of diabatic heating, as well as the mechanism of the diabatic heating effect on the intensity of the ZSL under the background of significant warming over the TP since the 1980s (Guo & Wang, 2011; Kuang & Jiao, 2016; Pang et al., 2020; B. Wang et al., 2008). Compared with previous studies whose data sets are of coarser temporal resolution, more detailed results can be revealed using the ERA5 reanalysis data sets with a finer resolution of an hour. This study focuses on the ZSL that induces heavy precipitation and is located in the ZSL high-frequency region over the TP. Moreover, rather than a case study, the composite analysis method is employed to draw more representative conclusions. The results can improve the understanding of the diabatic heating effect during the intensity evolution of the ZSL and provide some insight into the prediction of heavy precipitation over the TP.

This study is organized as follows: the data and methods are described in Section 2. Section 3 analyzes the diabatic heating effect on the intensity of the ZSL. Section 4 explores the mechanism of the diabatic heating effect on the intensity of the ZSL. Section 5 provides conclusions and discussions.

2. Data and Methods

As shown in Figure 1, the study region is 84°–96°E, 32°–35°N (orange-dashed box). The reasons for selecting this region are discussed below.

2.1. Data

The fifth generation ECMWF ERA5 hourly data sets (Hersbach et al., 2018) from June to August during 1980–2019 provided by the ECMWF are adopted in this study. The data has a horizontal resolution of 1° × 1° and 37 levels in the vertical direction. The 24-h accumulated rainfall collected from national meteorological stations, that is, the daily precipitation data sets (version 3.0), whose quality control is applied by the National Meteorological Information Centre of the China Meteorological Administration, was also used.
2.2. Methods

2.2.1. Objective Identification and Composite Methods for the ZSL

The objective identification method for the ZSL (Guan et al., 2018; Yao, Zhang, & Bao, 2020; Yao, Zhang, & Ma, 2020; X. Zhang et al., 2016; S. Zhang et al., 2019) is adopted in this study. The ZSL is identified by a combination of three parameters: the meridional shear of zonal wind, the vertical relative vorticity (hereafter referred to as vorticity), and the zonal wind velocity. Only when all three criteria in Equation 1 are met can a shear line be identified. Furthermore, given the scale of the ZSL, the identified shear line’s zonal span must be longer than 500 km (approximately longer than five degrees of longitude) to be identified as a ZSL.

\[
\begin{align*}
\frac{\partial u}{\partial y} &< 0 \\
\zeta &> 0 \\
u &> 0
\end{align*}
\]  

(1)

where \( u \) is the zonal wind velocity, \( v \) is the meridional wind velocity, \( y \) represents the coordinate in the meridional direction, and \( x \) represents the coordinate in the zonal direction. \( \zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \) is the vorticity, whose value is determined jointly by \( \frac{\partial v}{\partial x} \) and \( \frac{\partial u}{\partial y} \).

Using the objective identification method, 223 cases of ZSLs with life cycles of more than one day have been identified. These cases are further filtered, considering the ZSL’s birthplace, precipitation intensity, and life cycles, as shown by the following three requirements.

1. According to the geographic distribution, 97 cases situated in the high-frequency region (32°–35°N) of the ZSL (X. Zhang et al., 2016) are first selected.
2. With a focus on the cases that induced heavy precipitation around the ZSL, the standard deviation method is used, that is, if the precipitation of one day in the ZSL process exceeds the daily average precipitation of all the cases by more than one standard deviation, it can be regarded as a ZSL case. As a result, 43 out of the 97 cases of ZSL that meet this condition are selected second.
3. Based on the initiation and dissipation time of the ZSL, the life cycles of ZSL can be inferred. Therefore, 11 cases with life cycles longer than 60 h are finally selected out of the 43 cases.

Therefore, the 11 cases of ZSLs that meet the above requirements from June to August during 1980–2019 are selected and shown in Table 1. The local solar time (LST) is used for the time involved below. LST equals the coordinated universal time (UTC) plus 6 h, that is, LST = UTC + 6. It can be seen from Table 1 that most of the selected cases of ZSL initiated in the afternoon (14–18 LST) and dissipated during the morning hours (04–10 LST).

| Initiation time (LST) | Dissipation time (LST) | Life cycles (h) |
|-----------------------|------------------------|-----------------|
| 1980-07-05 16         | 1980-07-08 07          | 64              |
| 1982-06-24 17         | 1982-06-27 05          | 61              |
| 1983-06-26 15         | 1983-06-29 12          | 70              |
| 1985-06-07 15         | 1985-06-10 06          | 64              |
| 1985-08-21 14         | 1985-08-24 06          | 69              |
| 1987-07-08 16         | 1987-07-11 09          | 66              |
| 1991-08-04 14         | 1991-08-07 10          | 69              |
| 1992-06-22 17         | 1992-06-25 14          | 70              |
| 1992-06-25 17         | 1992-06-28 10          | 66              |
| 1996-07-23 14         | 1996-07-26 04          | 63              |
| 2018-06-11 18         | 2018-06-14 12          | 67              |

Note. For example, 1980-07-05 16 refers to 16 LST on July 5, 1980.

LST, Local solar time.

The value of vorticity is used to characterize the intensity of the ZSL. The arithmetic averages of the selected cases’ physical quantities are adopted for the composite analysis. The life cycle of the composited ZSL is 72 h, which initiates at 13 LST on the first day and dissipates at 12 LST on the fourth day. It covers all the processes of the initiation, maturity, and dissipation of the ZSL.

Based on the high-frequency region of the ZSL (X. Zhang et al., 2016) and the distribution region of the positive vorticity near the ZSL, the study region is selected as 84°–96°E, 32°–35°N (orange-dashed box in Figures 1 and 5). Subsequent calculations on regional average are all based on the study region.

2.2.2. Calculation of the Diabatic Heating

The “inverse algorithm” is used to calculate the diabatic heating over the TP, which is the most widely used calculation method at present (Lai & Gong, 2017). The following equation, based on the First Law of Thermodynamics, is used (Yanai & Johnson, 1993):

\[
\Delta H = \frac{\partial}{\partial y} (\rho u \frac{\partial v}{\partial x} - \rho v \frac{\partial u}{\partial y})
\]
In Equation 2, $Q_1$ is the diabatic heating over the TP; $C_p$ is the specific heat at constant pressure, which is 1004.8416 J kg$^{-1}$ K$^{-1}$; $T$ is the temperature; $V$ is the horizontal wind vector; $\omega$ is the vertical wind velocity in the pressure coordinate; $\kappa = 0.2875$; $\theta$ is the potential temperature and $p_0 = 1000$ hPa. The first term on the right side of the Equation 2 is the local variation of temperature. The second one is the horizontal advection term of temperature. The third one is the vertical transport term of temperature.

The vertically integrated diabatic heating $<Q_1>$ can be obtained using Equation 3.

$$<Q_1> = \frac{1}{g} \int_{p_s}^{p_t} Q dp$$

(3)

In Equation 3, $p_t$ is the pressure at the troposphere top (usually 100 hPa), and $p_s$ is the surface pressure.

The diabatic heating rate $Q$ can be obtained using Equation 4. The unit of $Q$ is K/s.

$$Q = Q_1/C_p = \frac{\partial T}{\partial t} + V \cdot \nabla T + \omega \left( \frac{p}{p_0} \right)^{\kappa} \frac{\partial \theta}{\partial p}$$

(4)

3. The Diabatic Heating Effect on the Intensity of the ZSL

The latent heating released by precipitation can enhance the diabatic heating and give positive feedback to the development of the ZSL. Therefore, it is crucial to study the diabatic heating effect on the intensity of the ZSL. In this section, the intensity evolution, the horizontal and vertical distribution, and the diabatic heating components around the ZSL during the ZSL life cycles are studied.

3.1. Intensity Evolution of the Diabatic Heating $<Q_1>$

Figure 2 shows the time series of the regionally averaged vorticity $\zeta$ at 500 hPa and the regionally averaged diabatic heating $<Q_1>$ (red line, unit: Wm$^{-2}$) and the regionally averaged vorticity $\zeta$ at 500 hPa (black line, unit: 10$^{-5}$ s$^{-1}$). The abscissa represents the time series, wherein solid black triangles represent the initiation and dissipation time of the zonal shear lie (the same below). The labels of 1st, 2nd, 3rd, and 4th represent the first day, the second day, the third day, and the fourth day, respectively (the same below).

As shown in Figure 2, the intensity variations of the ZSL and $<Q_1>$ show similar wave patterns, but one noteworthy finding is that the wave phase variation of $<Q_1>$ is 10 h earlier than that of the ZSL. Therefore, is there a correlation between the intensity variations of the ZSL and $<Q_1>$? Can the change of the intensity of $<Q_1>$ indicate the intensity evolution of the ZSL? It will be studied by the method of correlation analysis.

The correlation between the intensity of the ZSL and the intensity of the 10-h earlier regionally averaged $<Q_1>$ is shown in Figure 3. When the significance level $\alpha$ is set as 0.01, the correlation coefficient is as high as 0.81. Therefore, it can
be concluded that the relationship between them is significantly correlated at the significance level of \( \alpha = 0.01 \) (passing the 99% significance test).

As shown in Figure 4, the main body of the TP is a positive correlation zone between the vorticity at 500 hPa and the 10-h earlier regionally averaged \(<Q_1>\). The area with high positive correlations mainly appears near the ZSL and its southern side, with a maximum correlation coefficient exceeding 0.8. The significant positive-correlated large-value area near the ZSL indicates that the intensity variation of \(<Q_1>\) is closely related to the intensity evolution of the ZSL. The increase (decrease) of the intensity of \(<Q_1>\) may have an impact on the enhancement (weakening) of the ZSL.

### 3.2. Distribution Characteristics of the Diabatic Heating

As shown in Section 3.1, the intensity evolution of the ZSL is not synchronized with the \(<Q_1>\). Instead, a strong linear correlation shows a close relationship between the ZSL's intensity and the 10-h-earlier \(<Q_1>\). According to the life cycles of the ZSL and the intensity evolution of the vorticity, the first, second, and third days of the life cycles of the ZSL are defined as its initiation, maturity, and dissipation stages, respectively (From 13 LST on one day to 12 LST on the next day, with the daily average value representing the average situation of each stage). The horizontal and vertical distribution of the diabatic heating at the time before the initiation, maturity, and dissipation stages of the ZSL are analyzed to explore the effect of diabatic heating on the intensity evolution of the ZSL.

#### 3.2.1. Horizontal Distribution of the Diabatic Heating \(<Q_1>\)

The TP is dominated by the diabatic heating effect, showing a distribution characteristic of “high in the south and low in the north.” Because the study focuses on the diabatic heating effect in the intensity evolution of the ZSL, only the distribution characteristics of the \(<Q_1>\) in the study region (84°–96°E, 32°–35°N, the orange-dashed box) are discussed. The \(<Q_1>\) in the study region is more robust than that in other areas of the main body of the TP. The intensity of the \(<Q_1>\) at 10 h earlier than the initiation stage (Figure 5a) in the study region is about 300 Wm\(^{-2}\) and the maximum intensity of the ZSL is about \(4 \times 10^3\) s\(^{-1}\). At 10 h earlier than the maturity stage (Figure 5b), the range of the \(<Q_1>\) near the ZSL expands, and the maximum value exceeds 500 Wm\(^{-2}\). The intensity of the ZSL also increases subsequently. The maximum intensity of the ZSL exceeds \(4.5 \times 10^3\) s\(^{-1}\), and the range of large-value area is significantly expanded. At 10 h earlier than the dissipation stage (Figure 5c), the range of the \(<Q_1>\) near the ZSL changes little while the intensity decreases to 400 Wm\(^{-2}\). The intensity of the ZSL also weakens subsequently, and the range of large-value area is significantly reduced.

It can be seen from Figure 5 that the \(<Q_1>\) over the TP has a close relationship with the intensity evolution of the ZSL. When the \(<Q_1>\) over the southeastern side of the ZSL develops strongly, the ZSL subsequently strengthens. Therefore, the change in the intensity of the \(<Q_1>\) might be a good precursor of the intensity evolution of the ZSL.

#### 3.2.2. Vertical Distribution of the Diabatic Heating Rate \(Q\)

As shown in Figure 6, the diabatic heating rate \(Q\) varies by level, and the heating zone is mainly located above 500 hPa. The maximum value is at around 350 hPa, which may be related to the deep convection and latent heating released by condensation and deposition. The intensity of \(Q\) at 10 h earlier than the initiation stage (Figure 6a) over the ZSL is the weakest among all stages, with a maximum value of only 7 K day\(^{-1}\). At 10 h earlier than the maturity stage (Figure 6b), the intensity of \(Q\) near 350 hPa is rapidly enhanced to 10 K day\(^{-1}\). Combined with the evolutionary characteristics of the ZSL’s intensity, it is clear that the ZSL below the strong diabatic heating zone subsequently enhances. At 10 h earlier than the dissipation stage (Figure 6c), the intensity of \(Q\) gradually decreases. The distance between the ZSL and the diabatic heating center increases as the ZSL shifts northward. It results in weaker heating near the ZSL and a subsequent weakening of the intensity of the ZSL.
Figure 5. Horizontal distribution of the vorticity $\zeta$ at 500 hPa (contour, unit: $10^{-5} \text{s}^{-1}$) and the 10-h earlier $<Q_1>$ (shading, unit: W m$^{-2}$) at the (a) initiation stage, (b) maturity stage, and (c) dissipation stage. The orange-dashed box represents the study region (84°–96°E, 32°–35°N).
3.3. Components of the Diabatic Heating Rate $Q$

From Equation 4, it can be seen that the $Q$ obtained by the “inverse algorithm” is subject to the combined effect of the local variation of temperature term, the horizontal advection term of temperature, and the vertical transport term of temperature. During the intensity evolution of the ZSL, which one of the components contributes the most to the $Q$? What are the dynamical and thermal processes responsible for this? The corresponding conclusions will be given by exploring the contribution of each term on the right side of Equation 4.

Since the daily average value is used to represent the average situation of the stage, the local variation term of temperature in different stages is virtually unchanged. It contributes little to the $Q$. The $Q$ near and above 500 hPa is positive at 10 h earlier than the initiation stage (Figure 7a). The profile of $Q$ is very similar to that of the vertical transport term of temperature at the 500–300 hPa. Both of them reach the maximums at 350 hPa with intensities of about 6 K day$^{-1}$ and 7 K day$^{-1}$, respectively. Therefore, the vertical transport term of temperature makes a decisive positive contribution to the intensity of $Q$. Above 300 hPa, the positive contribution of the vertical transport term of temperature decreases, while the negative contribution of the horizontal advection term of temperature gradually increases. The combined effect of them makes the $Q$ tend to be around 0 K day$^{-1}$. Therefore, the diabatic heating effect on the upper levels is weaker than the lower levels.

The profile patterns at 10 h earlier than the maturity stage (Figure 7b) are similar to those of the initiation stage. The $Q$ and the vertical transport term of temperature still reach the peak value at 350 hPa, but the intensity is enhanced to 8 K day$^{-1}$ and 9 K day$^{-1}$, respectively. It indicates that the ascending motion near the ZSL is enhanced at the maturity stage. The intensity of the vertical transport term of temperature is slightly weakened at 10 h earlier than the dissipation stage (Figure 7c), and the height of the maximum value moves up to near 300 hPa. The $Q$ still reaches a maximum value at 350 hPa with little change in intensity. The horizontal advection term of temperature turns positive at heights from 350 hPa to near ground level, making $Q$ unchanged. However, the vertical transport term of temperature still plays a decisive role in the intensity of $Q$. It can also be found that the negative contribution of the horizontal advection term of temperature gradually increases from the initiation to the dissipation stage above 300 hPa, which may be related to the cooling above 300 hPa.

Figure 6. Vertical patterns of zonal-averaged 10-h earlier diabatic heating rate $Q$ (shading, unit: K day$^{-1}$), averaged over longitude 84°–96°E, at the (a) initiation stage, (b) maturity stage, and (c) dissipation stage. The black thick solid line denotes the zonal shear line (the same below). The gray shading is the terrain of the Tibetan Plateau (the same below).
In summary, the profile of \(Q\) is very similar to that of the vertical transport term of temperature at 500–300 hPa. The vertical transport of temperature is the main positive contributor to the intensity of \(Q\).

4. The Diabatic Heating Mechanism on the Intensity of the ZSL

From the previous analysis, it can be seen that the diabatic heating over the TP has a close relationship with the ZSL’s intensity. In addition to the horizontal advection term of vorticity and \(E\) effect, the diabatic heating effect is also taken into account in the complete-form vertical vorticity tendency equation (X. M. Wang et al., 2016; Wu & Liu, 1998, 1999). Therefore, the equation is used to explore the mechanism of the diabatic heating on the intensity evolution of the ZSL.

4.1. The Complete-Form Vertical Vorticity Tendency Equation

The complete-form vertical vorticity tendency equation (Wu & Liu, 1999) is as follows:

\[
\frac{\partial \zeta}{\partial t} + V \cdot \nabla \zeta + \beta \nu = (1 - \kappa) \left( f + \zeta \right) \frac{\partial \theta}{p} - \left( f + \zeta \right) \frac{Q}{\theta} + \frac{\partial}{\partial z} \left( \frac{\zeta}{\theta} \frac{\partial Q}{\partial \theta} \right) - \frac{1}{\theta} \frac{\partial}{\partial x} \left( \frac{\partial Q}{\partial x} \right) + \frac{1}{\theta} \frac{\partial}{\partial y} \left( \frac{\partial Q}{\partial y} \right) + R \tag{5}
\]

The terms on the right side of Equation 5 represent the effect of ascending motion, heat source, the spatially non-uniform diabatic heating (non-uniform diabatic heating is referred to as non-uniform heating, the same below), and the residual error \(R\), which consists of the frictional dissipation, the slantwise vorticity development, and computational error.) on the local variation of vorticity. \(p\) is the pressure; \(\theta\) is the potential temperature, and \(\theta_i = \frac{\partial \theta}{\partial z} \neq 0\). \(Q\) represents the diabatic heating rate in the thermodynamic equation, and Equation 4 is used to calculate it.

Equation 5 is used to explore the mechanism of the diabatic heating effect on the vorticity. Terms on the Equation 5 right side containing the effect of diabatic heating are the heat source term and the spatially non-uniform heating term. The magnitude of the heat source term is \(10^{-11} \text{ s}^{-2}\), which is about two orders of magnitude smaller than the spatially non-uniform heating term (10\(^{-9}\) s\(^{-2}\)). Therefore, this section focuses on the mechanism of the spatially non-uniform heating on the intensity evolution of the ZSL. The relationship between them is expressed as follows:
The spatially non-uniform heating term consists of the vertically non-uniform heating term and the horizontally non-uniform heating term, which characterizes the non-uniform state of the diabatic heating of the vertical and horizontal directions, respectively.

4.2. Spatially Non-Uniform Heating Effect

The vorticity variation and the spatially non-uniform heating effect near the ZSL are most significant at 500 hPa. It is clear from Figure 8 that the evolution of the local variation term of vorticity is consistent with that of the spatially non-uniform heating term.

It can be seen from Figures 9a–9c that the large value of the spatially non-uniform heating term is mainly concentrated below 400 hPa, and it always shows a heating effect near the ZSL. However, there are some differences in the heating intensity at different stages. At the initiation stage (Figure 9a), the intensity of the large-value center of the spatially non-uniform heating term is 1.6 × 10^9 s^{-2}. The maximum intensity of the ZSL increases to 1.9 × 10^9 s^{-2} at the maturity stage (Figure 9b). At the dissipation stage (Figure 9c), the maximum intensity of the spatially non-uniform heating term decreases to 1.6 × 10^9 s^{-2}, and the distance to the ZSL becomes longer. The intensity of the spatially non-uniform heating term near the ZSL weakens to about 0.6 × 10^9 s^{-2}.

In order to show the intensity variation of the spatially non-uniform heating term more clearly, we have added the analysis of the deviation, which is equal to the average of each stage minus the average of the entire life cycle. Figures 9d–9f show the deviation of the spatially non-uniform heating term in different stages. At the initiation and maturity stages (Figures 9d and 9e), there are positive deviations of the spatially non-uniform heating term near the ZSL. However, at the dissipation stage (Figure 9f), the deviation of the spatially non-uniform heating term turns negative, indicating a weakening of the heating effect near ZSL. These characteristics are consistent with the local variation term of vorticity near the ZSL, so the intensity variation of the spatially non-uniform heating term can influence the ZSL's intensity evolution.

4.2.1. Vertically Non-Uniform Heating Effect

In the northern hemisphere, it is known that $f = 2\Omega \sin \phi > 0$, and also $f + \zeta > 0$, $\theta = \frac{\partial \theta}{\partial z} > 0$ near the ZSL. Therefore, the positive and negative contributions of the vertically non-uniform heating term are determined by the vertical variation of the $Q \frac{\partial Q}{\partial z}$. If $\frac{\partial Q}{\partial z} > 0$, then $\frac{\partial \zeta}{\partial t} > 0$, and a positive vortex source is generated, which is beneficial to the development of vorticity; if $\frac{\partial Q}{\partial z} < 0$, then $\frac{\partial \zeta}{\partial t} < 0$, resulting in a reduction in the vorticity.

Figures 9a–9c and 10a–10c show that the distribution patterns of the spatially non-uniform heating term and the vertically non-uniform heating term are highly similar. The positive zones for both of them are concentrated below 400 hPa. Moreover, there is a large-value center at 500 hPa, corresponding to the typical level of the ZSL. Therefore, the vertically non-uniform heating term plays a decisive role in the distribution and evolution characteristics of the spatially non-uniform heating term. Figures 10d–10f show the deviation of the vertically non-uniform heating term in different stages. As similar to the spatially non-uniform heating term, there are positive deviations in the vertically non-uniform heating term near the ZSL at the initiation and maturity stages (Figures 10d and 10e). Whereas at the dissipation stage (Figure 10f), the deviation of the vertically non-uniform heating term turns negative, indicating a weakening of the heating effect near the ZSL.
Figure 9. Vertical patterns of zonal-averaged (a–c) spatially non-uniform heating term $f + \frac{\zeta}{\theta} \frac{\partial Q}{\partial z} + \frac{1}{\theta} \frac{\partial v}{\partial z} \frac{\partial Q}{\partial \theta} + \frac{1}{\theta} \frac{\partial u}{\partial z} \frac{\partial Q}{\partial \theta}$ and (d–f) its deviation (the average of each stage minus the average of the entire life cycle, unit: $10^{-3}$ s$^{-2}$), averaged over longitude 84°–96°E, at the (a and d) initiation stage, (b and e) maturity stage, and (c and f) dissipation stage.

To sum up, the vertically non-uniform heating effect in the vicinity of the ZSL is one of the primary mechanisms for the intensity evolution of the ZSL. When the vertically non-uniform heating effect near the ZSL enhances (weakens), it will be favorable (unfavorable) to enhance the ZSL.

4.2.2. Horizontally Non-Uniform Heating Effect

The magnitude of the horizontally non-uniform heating term ($10^{-10}$ s$^{-2}$) is one order of magnitude smaller than the vertically non-uniform heating term ($10^{-9}$ s$^{-2}$). The horizontally non-uniform heating term is composed of the meridionally non-uniform heating term and zonally non-uniform heating term, which characterizes the non-uniform state of the $Q$ in meridional and zonal directions, respectively.

The distribution and variation characteristics of the zonally non-uniform heating term (Figures 11d–11f) are consistent with those of the horizontally non-uniform heating term (Figures 11a–11c). Therefore, the horizontally non-uniform heating term is dominated by the zonally non-uniform heating term. However, its effect on the ZSL is not apparent and mainly affects the southern regions of the ZSL. A narrow zone of horizontally non-uniform heating effect exists near 32°N on the southern side of the ZSL in all three stages, and the center of the large value of heating is located near 500 hPa. It is most significant in the maturity stage (Figures 11b and 11e), which may be related to the abundant water vapor and precipitation occurring in this stage.
Figure 10. Vertical patterns of zonal-averaged (a–c) vertically non-uniform heating term $f + \zeta \frac{\partial Q}{\partial \theta}$ (unit: $10^{-7}$ s$^{-2}$) and (d–f) its deviation (the average of each stage minus the average of the entire life cycle, unit: $10^{-9}$ s$^{-2}$), averaged over longitude 84°–96°E, at the (a and d) initiation stage, (b and e) maturity stage, and (c and f) dissipation stage.

5. Conclusions and Discussion

Based on the ERA5 hourly reanalysis data with the spatial resolution of 1° × 1° from June to August during 1980–2019, 11 cases of heavy-precipitation-inducing ZSL with life cycles more than 60 h in the high-frequency region (32°–35°N) of the ZSL are selected. The mechanism of the diabatic heating effect on the intensity of the ZSL has been revealed by the methods of composite and diagnostic analysis. The main conclusions are as follows.

The intensity of the ZSL features a significant diurnal variation, with its peak at around 23 LST and its valley at around 15 LST within a day. The intensity of the vertically integrated diabatic heating $<Q_v>$ peaks at 13 LST within a day. Moreover, a close relationship exists between the ZSL’s intensity and the 10-h-earlier $<Q_v>$. The area with high positive correlations mainly appears near the ZSL and its southern side, with a maximum correlation coefficient of 0.81.

The diabatic heating rate $Q$ reaches its maximum value at around 350 hPa. The profile of $Q$ is very similar to that of the vertical transport term of temperature at 500–300 hPa. The vertical transport of temperature is the main positive contributor to the intensity of $Q$. 
The evolution of the local variation term of vorticity is consistent with that of the spatially non-uniform heating term. The spatially non-uniform heating term always shows a heating effect near the ZSL, while the heating intensity varies in different stages. The vertically non-uniform heating term plays a decisive role in the distribution and evolution characteristics of the spatially non-uniform heating term. Therefore, the vertically non-uniform heating effect in the vicinity of the ZSL is the primary mechanism for the intensity evolution of the ZSL. When the vertically non-uniform heating effect near the ZSL enhances (weakens), it will be favorable (unfavorable) to enhance the ZSL.

In this study, the relationship between the ZSL’s intensity and the intensity of diabatic heating has been revealed. The mechanism of the diabatic heating effect on the intensity of the ZSL has been obtained through the methods of composite and diagnostic analysis. The results have improved the understanding of the diabatic heating effect on the intensity evolution of ZSL and have provided some insight into the prediction of heavy precipitation over the TP. However, the conclusions obtained are based on a composite analysis of 11 cases of ZSL in a specific category. Further studies on the mechanism of the diabatic heating effect on the intensity of other types of ZSLs will be needed. Moreover, the effects of sensible heating, latent heating, and solar radiation on the evolution of the ZSL, respectively, need to be considered in the subsequent study.
The relationship between the ZSL and diabatic heating in terms of the ZSL lasting time and the direction of movement should also be examined in future research.

References

Abe, M., Hori, M., Yasunari, T., & Kitoh, A. (2013). Effects of the Tibetan Plateau on the onset of the summer monsoon in South Asia: The role of the air-sea interaction. *Journal of Geophysical Research: Atmospheres, 118*, 1760–1777. https://doi.org/10.1002/jgrd.50210

Chen, G., & Li, G. P. (2014). Dynamic and numerical study of waves in the Tibetan Plateau vortex. *Advances in Atmospheric Sciences, 31*(1), 131–138. https://doi.org/10.3878/j.issn.1006-9895.1906.18191

Ding, Y. H. (1992). Effects of the Qinghai-Xizang (Tibet) Plateau on the circulation features over the plateau and its surrounding areas. *Advances in Atmospheric Sciences, 9*(1), 112–130. https://doi.org/10.1007/BF02656935

Du, M., Li, G. P., & Li, S. S. (2020). A preliminary study of the relationship between the plateau transverse shear line and plateau vortex (in Chinese). *Chinese Journal of Atmospheric Sciences, 44*(2), 269–281. https://doi.org/10.3878/j.issn.1006-9895.1906.18191

Duan, A. M., Hu, D., Hu, W. T., & Zhang, P. (2020). Precursor effect of the Tibetan Plateau heating anomaly on the seasonal march of the East Asian summer monsoon precipitation. *Journal of Geophysical Research: Atmospheres, 125*, e2020JD032948. https://doi.org/10.1029/2020JD032948

Duan, A. M., & Wu, G. X. (2005). Role of the Tibetan Plateau thermal forcing in the summer climate patterns over subtropical Asia. *Climate Dynamics, 24*, 793–807. https://doi.org/10.1007/s00382-004-0488-8

Floh, H. (1957). Large-scale aspects of the “summer monsoon” in South and East Asia. *Journal of the Meteorological Society of Japan, 75*, 180–186. https://doi.org/10.2151/jmsj1923.35a.180

Guan, Q., Yao, X. P., Li, Q. P., Ma, Y. C., & Zhang, H. H. (2018). Study of a horizontal shear line over the Qinghai–Tibetan Plateau and the impact of diabatic heating on its evolution. *Journal of Meteorological Research, 32*(4), 612–626. https://doi.org/10.3878/j.issn.1006-9895.1906.18191

Guo, D. L., & Wang, H. J. (2011). The significant climate warming in the northern Tibetan Plateau and its possible causes. *International Journal of Climatology, 31*(12), 1775–1781. https://doi.org/10.1002/joc.2388

He, G. B., & Shi, R. (2011). Studies on dynamic and thermal characteristics of different shear line over Tibetan Plateau in summer (in Chinese). *Plateau Meteorology, 30*(3), 568–575.

Hersbach, H., Bell, B., Berrisford, P., Biavati, G., Horányi, A., Muñoz Sabater, J., et al. (2018). ERA5 reanalysis data sets (https://cds.climate.copernicus.eu) and the National Meteorological Information Centre for Medium-Range Weather Forecasts (ECMWF) for the ERA5 reanalysis data sets ([30x463](https://cds.climate.copernicus.eu) and the National Meteorological Information Centre for Medium-Range Weather Forecasts (ECMWF) for the ERA5 reanalysis data sets ([30x463](https://cds.climate.copernicus.eu))).

Kuang, X. X., & Jiao, J. J. (2016). Review on climate change on the Tibetan Plateau during the last half century. *Journal of Geophysical Research: Atmospheres, 121*, 3979–4007. https://doi.org/10.1002/2015JD024728

Lai, X., & Gong, Y. F. (2017). Relationship between atmospheric heat source over the Tibetan Plateau and precipitation in the Sichuan-Chongqing region during summer. *Journal of Meteorological Research, 31*(3), 555–566. https://doi.org/10.3878/j.issn.1006-9895.1906.18191

Li, L., Zhang, R. H., & Wen, M. (2011). Diagnostic analysis of the evolution mechanism for a vortex over the Tibetan Plateau in June 2008. *Advances in Atmospheric Sciences, 28*(4), 797–808. https://doi.org/10.3878/j.issn.1007-0102-0027-y

Li, S. S., & Li, G. P. (2017a). Diagnostic analysis based on wet Q-vector of a shear line with rain on the east side of Qinghai-Tibetan Plateau under the saddle pattern circulation background field (in Chinese). *Plateau Meteorology, 36*(2), 317–329.

Li, S. S., & Li, G. P. (2017b). Evolution and mechanism analysis of a plateau vortex and plateau shear line (in Chinese). *Chinese Journal of Atmospheric Sciences, 41*(4), 713–726.

Liu, G., Zhao, P., & Chen, J. M. (2017). Possible effect of the thermal condition of the Tibetan Plateau on the interannual variability of the summer Asian-Pacific Oscillation. *Journal of Climate, 30*, 9965–9977. https://doi.org/10.1175/jcli-d-17-0079.1

Liu, Y. M., & Wu, G. X. (2000). Reviews on the study of the subtropical anticyclone and new insights on some fundamental problems (in Chinese). *Acta Meteorologica Sinica, 58*(4), 500–512.

Liu, Y. M., Wu, G. X., Hong, J. L., Dong, B. W., Duan, A. M., Bao, Q., et al. (2012). Revisiting Asian monsoon formation and change associated with Tibetan Plateau forcing: I. Change. *Climate Dynamics, 39*, 1183–1195. https://doi.org/10.3878/j.issn.1006-9895.1906.18191

Luo, X., & Li, G. P. (2018). Numerical simulation and stage structure characteristics of a plateau shear line process (in Chinese). *Plateau Meteorology, 37*(2), 406–419.

Nan, S. L., Zhao, P., Yang, S., & Chen, J. M. (2009). Springtime tropospheric temperature over the Tibetan Plateau and evolutions of the tropical Pacific SST. *Journal of Geophysical Research, 114*, D10104. https://doi.org/10.1029/2008JD011559

Pang, H., Hou, S., Zhang, W., Wu, S., Jenk, T. M., Schwiokowski, M., & Jouzel, J. (2020). Temperature trends in the northwestern Tibetan Plateau constrained by ice core water isotope records across the past 7,000 years. *Journal of Geophysical Research: Atmospheres, 125*, e2020JD032560. https://doi.org/10.1029/2020JD032560

Qiu, J. (2008). China: The third pole. *Nature, 454*(7203), 393–396. https://doi.org/10.1038/454393a

The Tibetan Plateau Science Research Group. (1981). *Study on 500 hPa low vortex shear line of Tibet Plateau* (in Chinese). Beijing: Science Press.

Wang, B., Bao, Q., Hoskins, B., Wu, G. X., & Liu, Y. M. (2008). Tibetan Plateau warming and precipitation changes in East Asia. *Geophysical Research Letters, 35*, L14702. https://doi.org/10.1029/2008GL034330

Wang, X. M., Zhou, X. G., Tao, Z. Y., & Liu, H. (2016). Discussion on the complete-form vorticity equation and slantwise vorticity development. *Journal of Meteorological Research, 30*(1), 067–075. https://doi.org/10.3878/j.issn.1006-9895.1906.18191

Wu, D., Zhang, F. M., & Wang, C. H. (2018). Impacts of diabatic heating on the genesis and development of an inner Tibetan Plateau vortex. *Journal of Geophysical Research: Atmospheres, 123*, 11691–11704. https://doi.org/10.1029/2018JD029240

Wu, G. X., & Liu, H. Z. (1998). Vertical vorticity development owing to down-sloping at slantwise isentropic surface. *Dynamics of Atmospheres and Oceans, 27*, 715–743. https://doi.org/10.1016/s0377-0265(97)00040-7

Xie, H., & Liu, H. Z. (1999). Complete form of vertical vorticity tendency equation and slantwise vorticity development (in Chinese). *Acta Meteorologica Sinica, 57*, 1–14.

Wu, G. X., & Liu, Y. M. (2016). Impacts of the Tibetan Plateau on Asian climate. *Meteorological Monographs, 56*, 7.1–7.29. https://doi.org/10.1175/amsmonographs-d-15-0018.1

Xu, X. D., Lu, C. G., Shi, X. H., & Gao, S. T. (2008). World water tower: An atmospheric perspective. *Geophysical Research Letters, 35*(20), L20815. https://doi.org/10.1029/2008gl035867
Yanai, M., & Johnson, R. H. (1993). Impacts of cumulus convection on thermodynamic fields. In Emanuel, K. A., & Raymond, D. J. (Eds.), The representation of cumulus convection in numerical models. Meteorological Monographs. Boston, MA: American Meteorological Society. https://doi.org/10.1007/978-1-935704-13-3_4

Yanai, M., & Li, C. F. (1994). Mechanism of heating and the boundary layer over the Tibetan Plateau. Monthly Weather Review, 122(2), 305–323. https://doi.org/10.1175/1520-0493(1994)122<0305:mohatab>2.0.co;2

Yao, X. P., Sun, I. Y., Kang, L., & Ma, I. L. (2014). Advances on research of shear convergence line over Qinghai-Xizang Plateau (in Chinese). Plateau Meteorology, 33(1), 294–300.

Yao, X. P., Yan, L. Z., & Zhang, S. (2019). Research progresses and prospects of atmospheric diabatic heating (in Chinese). Meteorology Monthly, 45(1), 1–16.

Yao, X. P., Zhang, S., & Bao, X. H. (2020). Study on the structure of a horizontal shear line over the Tibetan Plateau based on CRA-Interim datasets and its comparison with ERA-Interim datasets. Journal of Tropical Meteorology, 26(4), 483–494. https://doi.org/10.46267/j.1006-8775.2020.042

Ye, D. Z. (1988). The thermal structure and the convective activity over Qinghai-Tibetan Plateau in summer and their interactions with large-scale circulation (in Chinese). Chinese Journal of Atmospheric Sciences, 12(s1), 1–12.

Ye, D. Z., Luo, S. W., & Zhu, B. Z. (1957). The wind structure and heat balance in the lower troposphere over Tibetan Plateau and its surrounding (in Chinese). Acta Meteorologica Sinica, 28, 108–121.

Zhang, S., Yao, X. P., & Gong, Y. F. (2019). A synthetic study of the structure and evolution characteristics of a meridionally-oriented shear-line over the Tibetan Plateau based on objective identification (in Chinese). Acta Meteorologica Sinica, 77(6), 1086–1106.

Zhao, P., Zhou, Z. J., & Liu, J. P. (2007). Variability of Tibetan spring snow and its associations with the hemispheric extratropical circulation and East Asian summer monsoon rainfall: An observational investigation. Journal of Climate, 20, 3942–3954. https://doi.org/10.1175/JCLI4205.1

Zhou, X. J., Zhao, P., Chen, J. M., Chen, L. X., & Li, W. L. (2009). Impacts of thermodynamic processes over the Tibetan Plateau on the Northern Hemispheric climate. Science in China Series D: Earth Sciences, 52(11), 1679–1693. https://doi.org/10.1007/s11430-009-0194-9