Interdecadal Variation of the Atmospheric Heat Source over the Tibetan Plateau and Surrounding Asian Monsoon Region: Impact on the Northern Hemisphere Summer Circulation

Xiaoting SUN\textsuperscript{1,2,3}, Yihui DING\textsuperscript{3}\textsuperscript{*}, and Qingquan LI\textsuperscript{1,3}

1 Key Laboratory of Meteorological Disaster, Ministry of Education/Joint International Research Laboratory of Climate and Environment Change/Collaborative Innovation Center on Forecast and Evaluation of Meteorological Disasters, Nanjing University of Information Science & Technology, Nanjing 210044
2 Chinese Academy of Meteorological Sciences, China Meteorological Administration, Beijing 100081
3 Laboratory for Climate Studies, National Climate Center, China Meteorological Administration, Beijing 100081

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ABSTRACT

We use 71-yr (1948–2018) reanalysis data to investigate the interdecadal variation in the atmospheric heat source ($Q_1$) over the Tibetan Plateau and surrounding Asian monsoon region (AMTP) and its effect on the Northern Hemisphere summer circulation. The large-scale circulation driven by $Q_1$ over the AMTP is characterized by a center of convergent (divergent) or low (high) potential wind function in the lower (upper) troposphere. $Q_1$ over the AMTP shows a clear interdecadal variation (with positive–negative–positive phases) and these three phases correspond to the time periods 1948–1972, 1973–2005, and 2006–2018, respectively. The thermal circulation has a corresponding interdecadal variation as a response to the interdecadal variation in $Q_1$. An enhanced $Q_1$ leads to an increase in the conversion of the total potential energy to non-divergent wind kinetic energy via the divergent wind velocity. The maximum conversion occurs in the tropopause. The primary thermal forcing for $Q_1$ is produced by the intense, large volume precipitation of the summer monsoon. This induces a response in the large-scale circulation, leading to large-scale divergence patterns. The synergistic effects of Pacific Decadal Oscillation (PDO) and North Atlantic Multidecadal Oscillation (AMO) influence $Q_1$ over the AMTP, which is ultimately responsible for the modulation of variations in the global divergent circulation. The global divergent circulation in summer is therefore essentially a direct thermodynamic circulation driven by the strong $Q_1$ over the AMTP.

Key words: Tibetan Plateau, atmospheric heat source, interdecadal variation, global divergent circulation, Asian–African summer monsoon

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1. Introduction

The Tibetan Plateau (TP) is the largest and highest plateau in the world. Ye and Gao (1979) reported that changes in the thermodynamic effects of the land–atmosphere system of the TP can affect the both East Asian summer monsoon (EASM) and the planetary-scale atmospheric circulation. Research on the role of the TP as a heat source started in the 1950s and showed that the surface of the plateau acts as a heat source throughout the year, whereas the atmosphere over the TP is a heat source in summer but a cold source in winter. The thermal condition of the TP has a great influence on the circulation of the atmosphere in summer and affects both the local-scale precipitation around the plateau and precipitation on larger scales. The strength of the TP atmospheric heat source ($Q_1$) affects the East Asian major trough, the summer South Asian high, and the strength of

\textsuperscript{*}Corresponding author: dingyh@cma.gov.cn

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the monsoons in both winter and summer (Ye and Gao, 1979; Tao and Ding, 1981; Murakami and Ding, 1982; Nitta, 1983; Luo and Yanai, 1984; Tao and Chen, 1987; Huang, 2006; Xu et al., 2013; Liu et al., 2018). Chen et al. (2007) and Krishnamurti et al. (2015) showed that there is a huge anticyclone near the tropopause over the TP and surrounding Asian monsoon region (AMTP) in summer, which forms in association with the large-scale heating field produced by vigorous precipitation in the Asian–African summer monsoon (AASM) system.

The recent increases in the area of snow cover in both winter and spring, the air temperature, and amount of precipitation over the TP have been shown to have a significant interdecadal variation (Zhu et al., 2007, 2009; Ding F. et al., 2008; Ding Y. H. et al., 2008). Wu et al. (2012a, b, 2016) found that snow cover on the western TP in summer is closely linked to the interannual variation in summer heat waves over southern Europe and northeastern Asia. There has been a significant change in the land–atmosphere coupling system over the TP with recent changes in the global climate, especially its thermodynamic component (Ding and Zhang, 2008). The surface temperature and precipitation over the TP have increased significantly in both winter and summer during the last two decades and show clear interdecadal changes with an increasing trend (Zhu et al., 2007, 2009; Duan et al., 2013; Si and Ding, 2013; Ding et al., 2015). The aim of this study is to demonstrate the interdecadal variations of $Q_1$ over the TP.

The AASM has an internally consistent variation on a planetary scale, which is generally assumed to have a synchronous interdecadal evolution (Ding and Li, 2016). The AASM began to strengthen after the end of the 20th century when the North Atlantic Multi-decadal Oscillation (AMO) entered a positive (warm water) phase. The interdecadal variability of the sea surface temperature (SST) in the North Atlantic and North Pacific oceans therefore affects the northward and southward migration of AASM precipitation (Li et al., 2017). Paleoclimate studies also indicate that weakening of the AASM is associated with cooling of the North Atlantic Ocean at high latitudes (Gupta et al., 2003; Shanahan et al., 2009; Stager et al., 2011). The synergistic effects of the Pacific Decadal Oscillation (PDO) and the AMO are the main reasons for the 30–40 yr periodic oscillation of the EASM, which has also undergone a strong–weak–strong interdecadal change since the 1960s, consistent with the trend of the AASM system (Ding et al., 2018). The TP is a subregion of the Asian summer monsoon (ASM) and the scientific question arises as to whether the anomalous interdecadal variability of the AMO and PDO is related to the anomalous $Q_1$ over the TP.

This paper focuses on the interdecadal variation of $Q_1$ over the AMTP in summer and the characteristics of the anomalous divergent wind fields in different interdecadal periods. The relationships between the interdecadal variation in $Q_1$, water vapor transport, and precipitation over the AMTP are discussed. The relationship between the PDO/AMO and the interdecadal variation in $Q_1$ is analyzed in terms of the forced influence of the North Atlantic and North Pacific oceans. The purpose of this study is to develop an understanding of the impact of the thermodynamic effects over the AMTP on the anomalous circulation in the Northern Hemisphere in summer and to provide a reference for future climate prediction studies.

2. Data and computational methods

We use the NCEP-1 monthly reanalysis dataset provided by the NCEP/NCAR from 1948 to 2018 (Kalnay et al., 1996). The data include the wind velocity ($V$), temperature ($T$), and geopotential height ($H$) from 1000 to 10 hPa. The vertical velocity data ($\omega$) include 12 layers from 1000 to 100 hPa and the specific humidity data ($q$) include 8 layers from 1000 to 300 hPa in the vertical direction. We use the sea-level pressure and a horizontal grid spacing of $2.5^\circ \times 2.5^\circ$.

We use the monthly NCEP-2 reanalysis data (Kanamitsu et al., 2002) from 1980 to 2018 and the Japanese 55-year Reanalysis (JRA-55) dataset (Kobayashi et al., 2015) provided by the Japan Meteorological Agency from 1979 to 2018 to calculate $Q_1$. These data include temperature, the vertical wind velocity, and the wind and sea-level pressure from 1000 to 100 hPa. The horizontal resolution of the NCEP-2 data is $1.875^\circ \times 1.875^\circ$ and that of the JRA-55 data is $1.25^\circ \times 1.25^\circ$. The monthly SST data [Extended Reconstructed Sea Surface Temperature version 4 (ERSST.v4)] from 1948 to 2018 were reconstructed by the NOAA (Huang et al., 2015). The monthly global precipitation data from 1948 to 2018 are provided by the NOAA Earth System Research Laboratory (Chen et al., 2002). The linear trends of the calculated $Q_1$ and other atmospheric circulation fields are removed to isolate the interdecadal variation of the atmospheric $Q_1$.

$Q_1$ is calculated by referring to the inverted algorithm of Yanai et al. (1992):

$$Q_1 = c_p \left[ \frac{\partial T}{\partial t} + V \cdot \nabla T + \left( \frac{p}{p_0} \right)^k \frac{\partial \theta}{\partial p} \right],$$  \hspace{1cm} (1)
where \( T \) is the temperature, \( \theta \) is the potential temperature, \( p \) is the pressure, and \( \omega \) is the vertical velocity. All these variables are in \( p \) coordinates. The variable \( p_0 (= 1000 \) hPa) is the standard pressure; \( k = R/c_p \approx 0.286 \), where \( R \) and \( c_p \) are the dry atmospheric constant and the isobaric specific heat capacity, respectively. \( V \) is the horizontal wind vector and therefore \( Q_1 \) can be calculated at each isobaric layer. From Eq. (1), \( Q_1 \) consists of three items. When calculating the atmospheric heat source \( \langle Q_1 \rangle \) in the troposphere, Yanai et al. (1992) set \( \omega = 0 \) at 100 hPa at the top of the troposphere. The vertical integral value of \( Q_1 \) is calculated as:

\[
\langle Q_1 \rangle = \frac{1}{g} \int_{p_s}^{p_t} Q_1 dp,
\]

where \( p_s \) is the surface level pressure and \( p_t (= 100 \) hPa) is the pressure at the top of troposphere. We select the nearest isobaric layer above the grid as the bottom level to integrate upward. Here, the range of the TP (28°–40°N, 70°–105°E), the ASM (25°–45°N, 60°–140°E), and southwestern North America (25°–35°N, 100°–110°W) are defined in latitude–longitude coordinates. Our calculation of \( Q_1 \) from 1948 to 2018 is consistent with the results of Ding et al. (2015).

According to the vorticity, divergence, and thermodynamic equations, the conversion function of the divergent wind kinetic energy (\( K_\psi \)) and the non-divergent wind kinetic energy (\( K_\chi \)) can be obtained with the following equation (Ding and Liu, 1985a, b; Ding et al., 1987):

\[
C(K_\psi, K_\phi) = f\nabla X \cdot \nabla \psi + \nabla^2 \psi \nabla X \cdot \nabla \psi + \frac{1}{2} \nabla^2 \chi + \omega J \left( \psi, \frac{\partial \chi}{\partial p} \right),
\]

where \( K_\chi \) is produced by \( P \), the potential energy. It is first computed with the baroclinic conversion function between \( P \) and \( K_\chi \):

\[
C(P, K_\chi) = -\nabla \cdot (\nabla \psi) + \nabla \cdot (\psi \nabla \chi) - \frac{\partial \psi}{\partial p} \cdot \frac{R}{P} \psi = \frac{\partial \psi}{\partial p} - \frac{R}{P} \psi.
\]

In Eqs. (3) and (4), \( \psi, \chi \), and \( \varphi \) represent the stream function, velocity potential, and potential height, respectively; \( f \) is the Coriolis parameter; and \( J \) and \( R \) are the Jacobian operator and specific gas constant, respectively. The four terms on the right-hand side of Eq. (3), i.e., \( f\nabla X \cdot \nabla \psi, \nabla^2 \psi \nabla X \cdot \nabla \psi, \frac{1}{2} \nabla^2 \chi \), and \( \omega J \left( \psi, \frac{\partial \chi}{\partial p} \right) \), are referred to as \( C_1 \), \( C_2 \), \( C_3 \), and \( C_4 \), respectively, in the following sections.

The physical understanding and computational procedure for Eqs. (3) and (4) are as follows. Initially, diabatic heating tends to change the available potential energy (\( P \)) through heating or cooling, and then, the available potential energy \( P \) changes the divergent wind \( K_\psi \) through Eq. (4). The changed \( K_\psi \) further transforms into \( K_\chi \) with Eq. (3). \( K_\chi \) is nearly equal to the actual or observed wind. These results are analyzed in Section 5.

### 3. Distribution of the global atmospheric heat source

Figure 1 shows the global distribution of \( Q_1 \) in summer (June–August) from 1948 to 2018. The distribution
of the vertically integrated $Q_1$ shows that the largest center is located in the ASM (Fig. 1). Large values of $Q_1 (> 200 \text{ W m}^{-2}$) are mainly located in the Bay of Bengal and over the southern slopes of the TP. This is consistent with the results of Yanai and Tomita (1998) and Zhang et al. (2012). In spring, autumn, and winter, the global centers of $Q_1$ are mainly located around the Intertropical Convergence Zone (figure omitted). In summer, the center near the equator moves to the ASM region as a result of the northward migration of the planetary-scale land–sea thermal contrast, forced by the northward shift in the latitude of solar declination. The TP is located in the strongest region of the global $Q_1$. However, the TP is not the center of the largest $Q_1$ as a result of the limited integration level (the atmosphere above 500–100 hPa) above the topography of the TP.

Figure 2 shows that the patterns of the potential function and divergent wind are similar to that of the water vapor flux in both the lower and upper troposphere. In general, there are two huge divergent wind centers on a global scale. One is located in the ASM and the other stretches from the equatorial eastern Pacific to the Atlantic oceans. The mid to lower troposphere (from the surface to 500 hPa) and the upper troposphere (400–100 hPa) in the ASM region present typical convergence and divergence patterns, respectively. The circulation pattern driven by $Q_1$ over the AMTP is mainly characterized by reverse circulation between the upper and lower layers—that is, large-scale wind convergence occurs in the mid to lower troposphere, whereas divergence occurs in the upper troposphere. The flow from the divergence center of the AMTP moves eastward, sinks over the tropical eastern Pacific Ocean and western coast of North America, and then flows back to the area of the Asian heat source with a low-level tropical east wind, thus forming a huge large-scale circulation cell on a global scale driven by the heat source in the ASM region.

Figures 2b and 2d indicate that the ASM region is the most powerful sink for water vapor convergence in the lower troposphere, whereas the eastern Pacific Ocean and equatorial Atlantic Ocean are sources of water vapor in the lower troposphere in summer. Previous studies have shown that the high water vapor content in the monsoon region, Central America, and equatorial West Africa is maintained by the divergent component of the water vapor flux, which is formed by the convergence of water vapor in these areas via a latitudinal Walker cell and a meridional Hadley cell (Salstein et al., 1980; Chen, 1985). Overall, the strongest regions of global $Q_1$, the convergence of low-level winds and the high-level cen-

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**Fig. 2.** Distributions of the potential function ($10^6 \text{ m}^2 \text{ s}^{-1}$) and divergence component ($\text{g s}^{-1} \text{ hPa}^{-1} \text{ cm}^{-1}$) of the (a, c) winds and (b, d) water vapor flux at (a, b) 300 hPa and (c, d) 850 hPa in summer from 1948 to 2018. The shading is the potential function of the wind field/water vapor flux, and the vectors are the convergence and divergence components of the potential function.
ters of divergence in summer are all located in the region of the AMTP.

There is an intense exchange of energy and water vapor between the TP and its surrounding atmosphere in the form of the sensible heat flux, latent heat flux, and radiation heating, which makes the TP a huge source of $Q_1$. Li and Yanai (1996) indicated that the rising flow caused by $Q_1$ over the TP and the convergence of the surrounding atmosphere favor the northward movement of tropical warm and humid air flows over the ASM. This process eventually leads to the onset of the EASM. Thermal and mechanical forcing by the TP leads to the first appearance of the ASM in the Bay of Bengal (Wu and Zhang, 1998) and then in the South China Sea (Ding and Liu, 2001). The variation of $Q_1$ over the TP in summer affects the distributions of wind convergence and divergence at different altitudes through forced heating. As a result, the global circulation and precipitation are significantly affected by the convergence and divergence caused by thermal imbalances.

We investigated the vertical profile of $Q_1$, the vertical velocity, and divergent winds along the latitude of the TP in summer from 1948 to 2018. Figure 3a clearly shows that the TP has the strongest $Q_1$ at this latitude. An intense $Q_1$ appears from the lower to upper troposphere. The center of $Q_1$ is located at 600–250 hPa, with a maximum of about $2.0 \times 10^{-5}$ K s$^{-1}$. There is also a large $Q_1$ over the Rocky Mountains in central and western North America, but its height is relatively low, located from the lower troposphere to 300 hPa. Another $Q_1$ near the east coast of North America is also high in the troposphere at 400–150 hPa. There is a large and shallow $Q_1$ near the surface in the deserts of North Africa, which is a result of the strong sensible heat flux in deserts in summer. From the vertical distribution of the divergent wind fields (Fig. 3b), the convergence occurs below 400 hPa over the TP. A divergence is located between 400 and 100 hPa in the troposphere with the central height at 150 hPa. The centers of convergence and divergence are located in the eastern TP. Another region of high wind potential is located in the North America–Atlantic Ocean–African desert region, which is opposite to its counterpart in the AMTP, with convergence in the upper layer and divergence in the lower layer. The updraft over the TP flows eastward to the eastern Pacific Ocean and the western coast of North America near 200 hPa and then changes into a downdraft. Overall, the TP is characterized by a strong $Q_1$ and upward motion in the troposphere, accompanied by low-level convergence and high-level divergence. This configuration of the circulation favors the development of convection over the TP in summer.

The profiles of $Q_1$ and vertical velocity over the TP are compared with those over southwestern North America (Figs. 4a, b). $Q_1$ is largest in the surface layer and decreases with increasing altitude; ascending motion dominates in the troposphere. The variations in $Q_1$ correspond with the variations in vertical velocity with a clear upward motion throughout the whole troposphere over the TP. Water vapor is transported from the southern TP and the ASM region in summer and the underlying surface is heated by the sensible heat flux. These thermal conditions favor the development of deep convection over the TP in summer.

The vertical profiles of $Q_1$ and the upward movement over southwestern North America are the opposite to those over the TP at all altitudes. The subsidence extends from the surface to the upper troposphere with a negative value of $Q_1$. A dynamic downdraft therefore prevails over this region, with cooling throughout the whole troposphere. Southwest North America is located on the southeastern side of the Pacific subtropical high. Weak subsidence prevails in the mid to upper troposphere as a result of large-scale convergence in this region in summer. Despite the larger sensible heat flux in the surface layer, the warming caused by subsidence restricts the thickness of convective clouds. The strong inversion caused by subsidence therefore does not favor the formation of deep convection and shallow convective clouds dominate in this region. From the viewpoint of the summer global circulation, the distributions of $Q_1$ and the vertical velocity in this area appear to be a dynamic compensation for thermal forcing over the AMTP.

### 4. Interdecadal variation of $Q_1$ over the TP

A high center of $Q_1$ in summer is located in the AMTP and drives the global-scale thermal circulation. A strong $Q_1$ and deep convection develop as a result of the transport of water vapor from the ASM region and thermal and dynamic uplift by the large-scale topography of the TP. The formations of $Q_1$ and convergence/divergence anomalies of large-scale circulation caused by $Q_1$ therefore have a close relationship with the topography of the TP.

#### 4.1 Temporal evolution of $Q_1$ over the TP

Figure 5 shows the summer $Q_1$ and its 10-yr low-pass filtering evolution over the TP and ASM regions. The long-term trend has been removed from the raw data series. The integrated values of $Q_1$ over the two regions show a similar interdecadal variation. $Q_1$ shows a clear positive–negative–positive interdecadal variation from 1948 to 2018. It then maintains a high value from the
1950s to 1970s before changing to a downward trend from the 1980s to the early 21st century, when it reaches a minimum value. $Q_1$ then shows an increasing trend and the intensity in the third phase is close to that in the first phase. We choose 1948–1972, 1973–2005, and 2006–2018 as the three phases of interdecadal variation in the intensity of $Q_1$ over the TP.

Figure 6 shows the time series and correlation coefficients of $Q_1$ over the TP derived from different reanalysis datasets. Despite the differences among the three time series, they all show relatively consistent trends on an interdecadal scale. $Q_1$ derived from the JRA-55 dataset represents a strong stage from the 1960s to 1970s. It clearly weakens from 1980 to 2000 and then increases gradually.

**Fig. 3.** Longitude–height cross-sections of (a) $Q_1$ ($10^{-5}$ K s$^{-1}$), (b) the wind potential function and divergence component ($10^6$ m$^2$ s$^{-1}$), and (c) the vertical velocity (Pa s$^{-1}$) averaged along 28°–40°N in summer from 1948 to 2018. In part (b), the shading is the wind field potential function averaged along 28°–40°N; the component of the vectors in the $x$-direction is the zonal wind $U$ (m s$^{-1}$) and the component of the vectors in the $y$-direction is $-\omega$ (hPa s$^{-1}$).
The NCEP-2 dataset also shows a weakening trend at the end of the 20th century with relatively small fluctuations. The correlation coefficient of \( Q_1 \) between the NCEP-1 and NCEP-2 datasets is 0.6933 from 1980 to 2018 and that between the NCEP-1 and JRA-55 datasets is 0.3852 from 1959 to 2018. Both correlations pass the 95% confidence level. Intercomparison of the three reanalysis datasets shows that the selected data reliably reflect the interdecadal variation of \( Q_1 \) over the TP, with generally consistent trends.

Figure 7 shows the vertical profiles of \( Q_1 \) in summer in different interdecadal phases. \( Q_1 \) is fairly large over...
the whole troposphere from the 1960s to 1970s, which implies that the convection is deeper. $Q_1$ shows a significant weakening trend from the 1980s to the beginning of the 21st century, especially in the mid to upper troposphere. Shallow convection is more frequent in the third period, corresponding to the enhancement of $Q_1$ in the lower and mid troposphere (700–400 hPa). The warmer and moister TP favors increased latent heating over this region. Overall, the value of $Q_1$ in different isobaric layers over the TP reflects interdecadal variations consistent with the integrated $Q_1$.

We calculated the 9-yr moving-average summer $Q_1$ over the TP and ASM at different isobaric surfaces (Fig. 8). The interdecadal variation in $Q_1$ over the TP presents consistent positive–negative–positive phases at different isobaric surfaces (600–100 hPa). These changes present more clearly in the upper troposphere than in the lower troposphere. The interdecadal variations in $Q_1$ over the ASM are similar to those over the TP, which generally present an increase from 1948 to 1972 and after 2005, whereas there is an anomalous negative phase during the 1970s–1990s in the mid to upper troposphere. This comparison led us to use a common name for the AMTP to characterize the features of its components of the TP and ASM sub-regions. Using the empirical orthogonal function from the ERA-40 data, Zhang et al. (2015) determined that the mid and late 1970s are the transition stage of the interdecadal variation in $Q_1$ in the ASM, which is roughly consistent with our study.

4.2 Large-scale divergent circulation in different interdecadal phases of $Q_1$ over the TP

The longitude–altitude profiles of $Q_1$, wind potential, and vertical velocity over the TP were calculated in three time periods. Figures 9a, 9d, and 9g show that $Q_1$ over the eastern TP increases significantly at 400–150 hPa during the first period (1948–1972). $Q_1$ over the western TP shows a positive anomaly from the surface to 100 hPa. The positive anomaly below 500 hPa may be dominated by the sensible heat flux. There is a small negative anomaly over the central plateau. By contrast, there are clear negative anomalies in the heat sources over the eastern and western TP when $Q_1$ weakens during 1973–2005. Combining the anomalous vertical velocity with the wind potential and its vertical circulation anomaly (Figs. 9b, e, h), the flow over most of the TP shows abnormal upward motion when $Q_1$ is increased. There is an abnormal convergent wind field from the lower layer to 150 hPa and divergence above 150 hPa. The maximum height of wind convergence occurs between 500 and 400 hPa in the first stage, whereas in the third stage, the maximum height occurs near 300–200 hPa. When $Q_1$ weakens in the second stage, the anomaly in the wind potential function shows divergence in the mid to lower troposphere and convergence in the upper troposphere.

Away from the TP, a negative $Q_1$ occurs on the west coast of North America and abnormal sinking below 200 hPa is observed during the strong $Q_1$ period. The vertical distributions of the wind potential and anomalous divergence occur from the surface to 300 hPa on west coast of North America. Anomalous convergence occurs in the upper troposphere. The trends of $Q_1$, vertical velocity, and wind potential in North Africa (part of the AASM regime) are consistent with those over the TP. Specifically, there is an abnormal upward motion in northern Africa.

![Fig. 7. Profiles of $Q_1$ (10^{-3} \text{ K s}^{-1})$ averaged over the TP in summer from 1948 to 2018.](image-url)
during the strong $Q_1$ period.

As a response to the interdecadal variability of $Q_1$ over the TP, the intensity of the divergence circulation on a global scale driven by the thermal forcing also shows corresponding interdecadal variations. Figure 10 shows the anomalous distribution of the global wind potential in three different periods and indicates that the divergence anomalies caused by thermal differences over the TP are clearly different at different stages. When $Q_1$ increases (Figs. 10a, b, e, f), the convergence circulation in the lower troposphere and the divergence circulation in the upper troposphere are strengthened over Eurasia and Africa. The center of divergence of the 100-hPa wind potential in the upper troposphere appears over the TP, Indian Peninsula, and Bay of Bengal, whereas the center of convergence appears in the eastern equatorial Pacific and on the western coast of South America. The flow on the TP is characterized by abnormal convergence at 850 Pa in the lower troposphere. The Sahel region of Africa and the northern Arabian Sea are also characterized by abnormal convergence, whereas the center of global divergence is located in the eastern equatorial Pacific and Central America. When $Q_1$ decreases over the TP (Figs. 10c, d), the anomalous convergence and divergence components show the opposite patterns.

These results show that the global divergent circulation in summer is essentially a direct thermal circulation driven by the strong $Q_1$ over the AMTP and this thermally driven circulation has clear interdecadal variations. The large-scale convergent wind fields over the AMTP can lead to dynamic sinking over the tropical eastern Pacific Ocean and the western coast of North America. Based on an analysis of the vertical motion and $Q_1$ profiles in Fig. 9, it can be shown that abnormal upward motion occurs from the near-surface to the mid to upper troposphere accompanied by an increase in $Q_1$. An
anomalous convergence is seen between 600 and 150 hPa and a divergence occurs above 150 hPa. By contrast, the global divergent circulation caused by thermal differences over the AMTP favors the further development of the Asian–African monsoon rain belt and the enhancement of convection via a feedback mechanism, thus maintaining a persistent interdecadal variation.

5. Conversion of energy between the divergent wind kinetic energy and the rotational wind kinetic energy forced by $Q_1$ over the AMTP

Chen and Wiin-Nielsen (1976) and Krishnamurti and Ramanathan (1982) showed that the total potential energy is first transformed into the kinetic energy of the divergent airflow. It is then transformed into the kinetic energy of the non-divergent airflow, which presents an increase in the actual wind field and the total kinetic energy. The divergent wind component therefore plays a key part in the development and evolution of the large-scale circulation (Xie et al., 1980; Ding and Liu, 1985a, b).

The vertical profiles of the total kinetic energy ($K$) over the TP are similar to that of the non-divergent wind kinetic energy ($K_{\psi}$) in different periods of $Q_1$ (Figs. 11a, c). The values of $K$ and $K_{\psi}$ in the troposphere increase during the enhancement period of the TP $Q_1$ and reach the maximum near 200 hPa. The maximum total kinetic energy over the TP was 9 J kg$^{-1}$ during 1948–1972 and 8 J kg$^{-1}$ during 2006–2018. However, when $Q_1$ weakened during 1973–2005, the intensities of $K$ and $K_{\psi}$ were weaker than those in the previous two periods. The variation in $K_{\psi}$ over the TP is consistent with $K$, which shows that the increase in the total kinetic energy mainly depends on the increase in the non-divergent wind kinetic energy. The difference in the divergent wind kinetic energy ($K_{\chi}$) appears at a higher altitude than the difference in $K$ and $K_{\psi}$, and its maximum value is near to 100–150 hPa at the tropopause. The value of $K_{\chi}$ increases in the periods (1948–1972 and 2006–2018) with a strong TP $Q_1$, especially in the third stage, but its amplitude is significantly smaller than $K$ and $K_{\psi}$. This shows that the ef-
effects of thermal differences on the divergent wind kinetic energy over the TP vary in different stages of $Q_1$. The characteristics of the $K_x$, $K_{x'}$, and $K_\psi$ profiles over the ASM (Figs. 11d–f) are very similar to those over the TP, but the area-average value is slightly weaker than that over the plateau.

Figure 12 shows the divergent wind kinetic energy and non-divergent wind kinetic energy conversion function $C(K_x', K_\psi)$ and its four subitems $C_1$, $C_2$, $C_3$, and $C_4$ in each pressure layer in different periods of $Q_1$ over the TP and ASM. The conversion function increases significantly from the tropopause above 150 hPa to the lower stratosphere at 50 hPa when the TP $Q_1$ is relatively strong (1948–1972). The maximum value is $2 \times 10^{-5}$ W m$^{-2}$ at 100 hPa and much of $K_x'$ is converted into $K_\psi$. At the same time, the negative $C(K_x', K_\psi)$ conversion increases in the mid and lower troposphere, which destroys $K_\psi$ at 500–150 hPa. During 2006–2018, the $C(K_x', K_\psi)$ profile is similar to that of the first stage, but the maximum height significantly increases. The negative conversion occurs from 300 hPa to the tropopause, whereas the conversion of $K_x'$ to $K_\psi$ is mainly manifested in the stratosphere (70–50 hPa). When there is a weak $Q_1$ (1973–2005), the conversion $C(K_x', K_\psi)$ is negative from 200 hPa to the lower stratosphere and relatively weak in the lower pressure layers. The characteristics of the transition between the ASM and TP are similar, but the ASM regional intensity is weak. The evolution of $C(K_x', K_\psi)$ shows that the kinetic energy conversion from $K_x'$ to $K_\psi$ over the ASM and TP mainly occurs from the tropopause to the lower stratosphere. $C_1$ is the most important of the four conversions, which indicates that relative ori-
entation of the gradient between the $\chi$ and $\psi$ fields has a decisive role in the conversion.

Figure 13 shows that the conversion function $C(P, K_\chi)$ of the vertical profile from the available potential energy to the divergent wind kinetic energy over the TP and ASM are consistent in different TP periods and the extreme values appear between the tropopause and lower stratosphere (about 100 hPa). When the diabatic heating over the TP or ASM region is enhanced during 1948–1972 and 2006–2018, the total potential energy increases and the conversion from the available potential energy to the divergent wind kinetic energy increases. Compared with 1948–1972, $C(P, K_\chi)$ increased over the TP during 2006–2018, with maxima of $24 \times 10^{-5}$ and $10 \times 10^{-5}$ W m$^{-2}$, respectively. When $Q_1$ over the TP is weak, the minimum is $-15 \times 10^{-5}$ W m$^{-2}$. The $C(P, K_\chi)$ vertical profile over the ASM region is consistent with the TP, but its maxima are lower than that of the TP in the same pressure layer.

6. Interdecadal variation in $Q_1$ over the AMTP region and its impact on circulation in the Northern Hemisphere

6.1 Characteristics of precipitation and water vapor transport

Figure 14 shows the regression coefficients of the global precipitation and the water vapor transport flux at 850 hPa against the summer $Q_1$ over the TP and ASM re-
regions from 1948 to 2018. The patterns of precipitation and water vapor associated with $Q_1$ over these two regions are fairly consistent. The precipitation over the southern TP clearly increases with increasing $Q_1$. The intensity in $Q_1$ is significantly correlated with the location of the AASM rain belt. The precipitation belt increases uniformly during the strong $Q_1$ period over the TP and ASM regions—that is, the precipitation increases in the Sahel, northwest India, northern China, and southern Japan. Previous studies (Ding and Li, 2016; Li et al., 2017) have indicated that the AASM has presented a consistent northward and southward evolution in Africa and Asia for the past 100 years. The planetary-scale AASM is an internally consistent monsoon system.

The AASM feedback mechanism mainly enhances the associated water vapor transport. The regression coefficients between $Q_1$ over the TP and the water vapor transport flux suggest that the water vapor is mainly transported from the southern slopes of the TP. The water vapor in the southwest path flows from western and central Africa to the TP through the Arabian Sea and the northwestern Indian Peninsula. The southeast path flows westward from the equatorial Pacific to the oceanic continent. It then turns northwestward to the southeastern TP. The
water vapor is lifted up on the southern slopes of the TP, which leads to an increase in latent heating and $Q_1$.

The regression between $Q_1$ over the TP (Figs. 15a, c) and the global vertical velocity at 200 and 600 hPa shows that the southern TP has an abnormal upward motion with increasing $Q_1$. An abnormal upward motion also occurs in North–Central Africa, northwestern Indian Peninsula, southern TP, and northern China. Dominant downward motion occurs in southern China and near the Philippines. The vertical velocity pattern of $Q_1$ (Figs. 15b, d) associated with the ASM region has a similar distribution, showing a broader and stronger upward movement from the mid and lower troposphere to upper layers. At the same time, the AASM also shows a consistent abnormal upward movement. $Q_1$ is closely related to water vapor convergence and the ascending motion of the AASM. Combined with the regression of the 300–500-hPa vertical velocity (figure omitted), the distribution characteristics are consistent with those at 200 and 600 hPa.

6.2 Global teleconnection path enhanced by $Q_1$ over the AMTP region in summer

Figure 16 shows the regression of the geopotential heights at 200 and 600 hPa against $Q_1$ over the AMTP in summer. The lower and midtroposphere over the TP correspond to an abnormally low pressure when $Q_1$ increases, whereas an abnormally high pressure appears in the upper troposphere. An enhanced $Q_1$ corresponds to upward motion, which favors the formation of a thermal depression and low-level convergence over the TP, while the upper level divergence is strengthened. By contrast, the eastern Pacific Ocean and the west coast of North America are an anomalous high-pressure region. As a response to thermal forcing, the geopotential height shows an east–west-oriented positive–negative–positive pattern in the troposphere at mid to high latitudes, with multiple centers located in the Okhotsk Sea, Alaska, North America, and near southeastern North America. The geopoten-
Potential height over the Okhotsk Sea and Greenland is positively correlated with $Q_1$, whereas that over Alaska and southeastern North America is negatively correlated with $Q_1$. There are abnormal low- and high-pressure regions in western Europe and the Ural Mountains, respectively. The potential height distribution is reversed when the TP $Q_1$ weakens.

These results show that the increase in $Q_1$ affects the wave intensity at mid to high latitudes along the global teleconnection path originating from the North Atlantic Ocean in the Northern Hemisphere (Li et al., 2017). The vertical regression coefficients of the geopotential height
within 28°–40°N and 50°–70°N against \( Q_1 \) over the TP are calculated (Fig. 17). The centers of the geopotential height anomaly in the Okhotsk Sea, Alaska, and northern North America always show a barotropic structure characterized by a consistent positive–negative–positive pattern from the lower to upper troposphere (Fig. 17a). The anomalous center of geopotential height over the TP is negative below 400 hPa and positive above 400 hPa, which indicates a baroclinic structure mainly affected by thermal forcing (Fig. 17b). The regression coefficients over the eastern Pacific and on the west coast of North America show a roughly barotropic structure related to the sinking of large-scale convergent winds caused by thermal forcing over the TP.

7. Discussion and conclusions

7.1 Discussion

The PDO and AMO are two interdecadal oscillation modes of the SST. These two interdecadal oscillation modes are the main natural driving forces of the ASM (Zhu and Yang, 2003; Li et al., 2017; Ding et al., 2018). Changes in the PDO and AMO occur over long time periods and show a periodicity in phase. Forced coupling of these oscillation modes exerts a significant impact on the interdecadal variation of the ASM.

Figure 18 shows the regression coefficients between the AMTP \( Q_1 \) and global SSTs. The increase in the SST in mid to high latitudes over the western and central Pacific Ocean and the decrease in the eastern Pacific Ocean are closely related to the increase in the \( Q_1 \) anomaly. This distribution is consistent with a negative (cold) PDO phase. By contrast, anomalous high SSTs occur over the North Atlantic when \( Q_1 \) increases, showing a positive (warm) AMO phase.

Combined with the AMO index (Trenberth and Shea, 2006) and the PDO index (Mantua et al., 1997) from 1948 to 2018 (Fig. 19), the mean values of \( Q_1 \), PDO, and AMO in three interdecadal phases are calculated. The PDO index presents a negative–positive–negative trend on an interdecadal scale, whereas the trend of the AMO index is the opposite. A cold phase of the PDO persisted during the periods with a strong \( Q_1 \) (1948–1972 and 2006–2018), whereas the PDO index was in a warm

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**Fig. 17.** Profiles of the regression coefficients of the geopotential height \([10^{-3} \text{ gpm} (\text{W m}^{-2})^{-1}]\) averaged along (a) 50°–70°N and (b) 28°–40°N and \( Q_1 \) averaged over the TP in summer from 1948 to 2018. The black dots represent grid points with statistical significance exceeding the 90% confidence level.
phase in 1973–2005 when a weak $Q_1$ occurred. By contrast, the AMO index showed a positive anomaly (warm phase) in time periods with a strong $Q_1$. The AMO index decreased and became a negative anomaly (cold phase) in time periods with a weak $Q_1$. The intensity of $Q_1$ is therefore negatively correlated with the PDO index and positively correlated with the AMO index.

Our results show that, to a large extent, the interdecadal variation in $Q_1$ over the AMTP is forced by coupling of the AMO and PDO indexes (AMO$^+$/PDO$^-$ and AMO$^-$/PDO$^+$). The intensity of $Q_1$ will increase when the AMO is in a warm phase and the PDO is in a cold phase, and vice versa. Previous studies have shown that the PDO may be the main factor driving the first mode of summer precipitation in East Asia and is inversely correlated with precipitation in the Yangtze–Huaihe River basin (Ding et al., 2018). The AMO can influence interdecadal variations in precipitation over entire Northern Hemisphere from the Atlantic Ocean, Eurasia to North America by exciting a global-scale baroclinic teleconnection (Si and Ding, 2016; Li et al., 2017). The studies on EASM precipitation during different PDO and AMO phases indicate that the SST in the North Pacific and North Atlantic oceans shows the same-sign anomalies as the monsoon precipitation when the PDO and AMO indexes are out of phase (Zhang et al., 2018). At the same time, a zonal teleconnection wave train is seen at mid to high latitudes in Eurasia. This wave train propagates along the Asian westerly jet waveguide from the North Atlantic Ocean to Northeast Asia, forming a global zonal Rossby wave train spanning the mid to high latitudes of Eurasia from the North Atlantic to North Pacific oceans. This wave train connects the AMO-related North Atlantic–European east–west mode and the PDO-related North Pacific barotropic atmospheric circulation, which both affect the rain belt of the EASM region.

7.2 Conclusions

We analyzed the interdecadal variation in $Q_1$ over the AMTP in summer and showed the characteristics of the divergence wind fields and vertical velocity anomalies in different interdecadal periods. We also analyzed their effects on the atmospheric circulation and precipitation of the Northern Hemisphere in summer.
One key finding is that the TP is located in the AMTP region, which has the strongest $Q_1$ in summer. The circulation driven by the huge $Q_1$ over the AMTP is mainly represented by the centers of divergent winds in the upper and lower layers. Large-scale wind convergence and divergence occur in the mid to lower and upper troposphere, respectively. An air current from the center of divergence over the AMTP flows from west to east and sinks over the tropical eastern Pacific Ocean and the western coast of North America. It then flows back to the AMTP region with a low-level tropical east wind, forming a large-scale circulation cell driven by thermodynamic effects on a global scale.

The variable $Q_1$ over the AMTP has clear interdecadal variations. During 1948–1972, $Q_1$ maintained a high intensity in summer. However, it showed an anomalous negative phase from 1973 to 2005 and strengthened again from 2006 to 2018. The diagnostic computation with the $\chi-\psi$ energy conversion method showed that when $Q_1$ over the AMTP is enhanced, the conversions from total potential energy to divergent wind kinetic energy and then to non-divergent wind kinetic energy are clearly increased. As a response to the interdecadal variation in $Q_1$, this global circulation also has a corresponding interdecadal variation. An anomalous upward motion occurs in the mid and upper troposphere when $Q_1$ over the AMTP is strengthened. The wind fields from low levels to 150 hPa and above 150 hPa show anomalous convergence and divergence, respectively. There is an abnormal compensatory subsidence of convergence over the tropical eastern Pacific Ocean and west coast of North America. These results show that the large-scale divergence caused by thermal forcing over the AMTP can cause an anomalous compensatory dynamic sinking in remote areas.

The intensity of $Q_1$ over the TP is significantly correlated with the AASM rain belt. During the high-intensity period of the AASM, the northward movement of the rain belt favors the release of latent heat over the TP, and $Q_1$ increases. The interdecadal variation in $Q_1$ and the intensity of the ASM are ultimately forced by coupling of the AMO and PDO, as documented by this work and other studies (e.g., Li et al., 2017). $Q_1$ is strengthened during the warm phase of AMO and the cold phase of PDO, and vice versa. The natural forcing of interdecadal coupling in the oceans is ultimately responsible for the global divergent circulation by affecting $Q_1$ over the AMTP.

The interdecadal oscillation of the SST over the Pacific and Atlantic oceans in summer modulates the global divergence circulation through affecting $Q_1$ over the TP or AMTP. The thermal circulation as a thermal forcing generates dynamic forcing over the tropical eastern Pacific Ocean and the west coast of North America. This is a consequence of the natural driving force. We have not discussed the effect of anthropogenic forcing in this paper. It should be pointed out that the role of human factors will gradually increase with ongoing global warming. It is worth investigating when the effect of human activities on the global circulation will be close to or exceed the natural variability of the climate system. We plan to use multiple numerical models to analyze this issue further.

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