Impacts of the Indian Summer Monsoon on the southern boundary water vapor transport and precipitation over the Tibetan Plateau

Tianyu Liu
Nanjing University of Information Science and Technology

Jinghua Chen (jhchen@nuist.edu.cn)
Nanjing University of Information Science and Technology

Kai Yang
Nanjing University of Information Science and Technology

Liping Deng
Guangdong Ocean University

Research Article

Keywords: Tibetan Plateau, Water vapor transport, India summer monsoon, Precipitation

Posted Date: September 15th, 2022

DOI: https://doi.org/10.21203/rs.3.rs-2046021/v1

License: © This work is licensed under a Creative Commons Attribution 4.0 International License. Read Full License
Impacts of the Indian Summer Monsoon on the southern boundary
water vapor transport and precipitation over the Tibetan Plateau

Tianyu Liu\textsuperscript{1,2}, Jinghua Chen \textsuperscript{1,2}, Kai Yang\textsuperscript{1,2}, Liping Deng\textsuperscript{3}

\textsuperscript{1}Key Laboratory for Aerosol-Cloud-Precipitation of China Meteorological
Administration, School of Atmospheric Physics, Nanjing University of Information
Science & Technology, Nanjing 210044, China

\textsuperscript{2}Collaborative Innovation Center on Forecast and Evaluation of Meteorological
Disasters (CIC-FEMD), Nanjing University of Information Science & Technology,
Nanjing, China

\textsuperscript{3}College of Ocean and Meteorology Guangdong Ocean University Zhanjiang,
Guangdong China, 524088

Corresponding author: Jinghua Chen, jhchen@nuist.edu.cn
Abstract

Water vapor transport plays a significant role in maintaining the water cycle over the Tibetan Plateau (TP). This study investigates the characteristics of water vapor transport across TP southern boundaries and its impacts on TP precipitation during the Indian summer monsoon (ISM) season from 2000 to 2019. The southern boundary is subdivided into four sub-boundaries (boundaries 7, 8, 9, and 10) from the east to the west. Water vapor transports of boundaries 7, 8, and 9 are mainly affected by ISM, while mid-latitude westerlies dominate the water vapor transport of boundary 10. The results show that the PCR precipitation concentrated over the center TP in both ISM strong and weak months is smaller in the ISM weak months than the normal months for most of the day, while it is larger in the ISM strong months than the normal months. The PCR precipitation correlates positively with the water vapor transport across boundary 10 in both the ISM strong and weak months. Although there is water vapor transport across boundary 7 in both the ISM weak and strong months, the water vapor can hardly be brought to PCR. The correlation between the PCR precipitation and the water vapor from the BOB and the Arabian Sea is more intense in ISM strong months than in ISM weak months. Conversely, the water vapor transport efficiency is low in ISM strong months due to a cyclonic circulation over northern India, preventing water vapor transport from reaching the TP directly.

Keywords Tibetan Plateau · Water vapor transport · India summer monsoon · Precipitation

1 Introduction

The Tibetan Plateau (TP), known as "the world water tower", is an essential water tower in the world for two reasons (Xu et al. 2008). Firstly, the TP is the highest land in the world with an average altitude of over 4,000 m and is referred to as the "roof of the world". Secondly,
a large amount of water is stored and maintained in the TP in all its forms, as it includes more
than $1 \times 10^4$ km$^2$ of glaciers (Qiu 2008; Yao et al. 2012), over $5 \times 10^4$ km$^2$ of lakes (Li et al.
2014), and approximately $1.06 \times 10^6$ km$^2$ of permafrost (Zhao et al. 2020). Meanwhile, the TP
is also the origination of more than ten major rivers, such as the Yangtze, the Yellow River, and
the Mekong River (Li et al. 2015; Yao et al. 2012). Abundant water resources are stored in the
TP, while a large amount of water is flowing out of the TP at the same time. That means the
maintenance of the regional water cycle in the TP needs water sources outside the TP, and an
important source is the water vapor transported into the TP by the circulation (Bothe et al. 2012;
plays a significant role in maintaining the water cycle over the TP, and this topic has invoked
the interest of many researchers (Sun et al. 2022; Xu et al. 2020; Yu et al. 2022; Zhou et al.
2019).

For the TP, in addition to the evaporation from the local surface, moisture from outside,
e.g., the Bay of Bengal (BoB), the Arabian Sea, and the mid-latitude westerlies, is the most
critical water vapor source in summer (Bothe et al. 2012; Chen et al. 2012; Wang et al. 2017;
Li et al. 2019). The TP locates at the convergence of the mid-latitude westerlies and the South
The large-scale circulations provide favorable dynamic conditions for water vapor transport to
the TP. Previous studies have noted that water vapor enters the TP mainly across two
boundaries (Feng and Zhou 2012; Simmond et al. 1999; Xu et al. 2020; Zhu et al. 2015). One
is the western boundary dominated by mid-latitude westerlies. Another one is the southern
boundary, which is mainly influenced by the Indian summer monsoon (ISM). The southern
boundary has been identified as the boundary with the most considerable water vapor input (Feng and Zhou 2012; Simmonds et al. 1999; Xie et al. 2019). The water vapor transported by the ISM from the Arabian Sea and the BoB across the southern boundary accounts for over 60% of total transported moisture into the TP (Lin et al. 2016).

The southern boundary of the TP is a complex terrain with the highest mountains in the world, the Himalayas. On the one hand, mountains form a mechanical barrier to the water vapor input (Lu et al. 2005), part of the water vapor flow is blocked by the plateau and then deflected predominantly to the east side of it after reaching the TP (Xu et al. 2010). On the other hand, there is intense surface sensible heating (SH) appears over the southern slope of the TP in the summer (Wang et al. 2016; Wu et al. 2007). Xu et al. (2014) revealed that water vapor could be transported to the TP through a CISK (conditional instability of the second kind)-like mechanism. The warm–moist airflow rises to the plateau and then transports to the center of it, forced by the surface heating, resulting in frequent convection and precipitations over the TP (Xu et al. 2003).

Since water vapor is an indispensable material for the formation of precipitation, many previous studies (Dong et al. 2016; Feng and Zhou 2012; Xu et al. 2020; Zhang et al. 2019) have emphasized the significance of the southern boundary as a vital water vapor input boundary for the precipitation of the plateau in summer. Feng and Zhou (2012) addressed that water vapor transport from the southern boundary dominates the summer precipitation over the southeastern TP. This point is supported by Zhang et al. (2019), which estimated that more than half of the precipitation moisture in southern TP is transported from the southwest of the TP with the breakout of the Indian monsoon in summer. Dong et al. (2016) indicated water vapor
transport to the TP via the up-and-over route. In this route, moisture is lifted by convective
storms along the south slope of the plateau and then swept over the southwestern TP (SWTP)
by the mid-tropospheric circulation. Moisture transported in this pathway accounts for over half
of total summer rainfall over the SWTP. Therefore, this study focuses on the water vapor
transport across the southern boundary and its impact on the precipitation in the TP.

Although there are many previous studies on water vapor and water vapor transport over
the TP, most of them as mentioned above are mainly based on monthly or longer time scales.
Few studies focus on the daily and sub-daily variations of water vapor transported into the TP.
Moreover, the detailed characteristics of water vapor transport across southern boundaries and
its impact on the precipitation in the TP still need to be further investigated. This paper analyses
the characteristics of water vapor transport across the southern boundary and the corresponding
precipitation characteristics over the TP in the monsoon season (June through September).
Possible responsible for the precipitation change in the TP are further discussed.

The rest of this paper is organized as follows. The data and method used in this study are
shown in Section 2. Section 3 shows the characteristics of water vapor transport and
precipitation over the TP under the influence of ISM. Finally, the conclusions and discussion
are given in Section 4.

2 Data and Method

The ERA5 hourly reanalysis data for the duration of 2000–2019 in the monsoon season
(June through September) produced by The European Centre for Medium-Range Weather
Forecasts (ECMWF) is used in this study. Compared with ERA-Interim, ERA5 has a higher
spatial and temporal resolution (Hoffmann et al. 2019; Lauril et al. 2021), so it performs better
in capturing the spatial pattern of water vapor transport over the TP (Jiang et al. 2021; Slättberg et al. 2022; Sun et al. 2022; Zhao and Zhou 2020). To analyze the water vapor transport to the TP, the hourly specific humidity, temperature, surface latent heat flux, surface sensible heat flux, surface pressure, total precipitation, and wind field from the ERA5 reanalysis data set are used in this study, which has a horizontal resolution of $0.25^\circ \times 0.25^\circ$ with 137 vertical levels (Hersbach et al. 2020).

The hourly CMORPH precipitation product, which is merged by China Automatic Weather Station (CAWS), is used in this study. The time period is 2000-2019 in the monsoon season (June through September). The probability density function–optimal interpolation (PDF-OI) methods have been applied in this product to merge the gauge observations from more than 30,000 automatic weather stations in China with the Climate Precipitation Center Morphing (CMORPH) precipitation product (Shen et al. 2014). The improved merged precipitation product over China at hourly, 0.1° resolution has been certified as having better quality than the CMORPH and is more applicable to the precipitation study in China (Pan et al. 2012).

The Asian summer monsoon (ASM) is the largest and strongest monsoon system in the world (Liu et al. 2019), and the Indian summer monsoon (ISM) is one of the monsoon subsystems of ASM. The strength of ISM can be characterized by the ISM Index (IMI). Wang and Fan (1999) proposed an IMI, which is defined as the difference of the 850-hPa zonal winds between a region in the Arabian Sea (5-15° N, 40-80° E) and a region in eastern TP (20-30° N, 70-90° E) and this index is used in this study. The zonal wind shear has a stronger correlation with the rainfall, and this IMI represents well the rainfall anomalies averaged over the BoB, India, and the eastern Arabian Sea (Wang et al. 2001).
The ISM strong and weak months are classified based on the standardized anomaly of IMI in the monsoon season (June through September) for the years 2000-2019. An ISM strong (weak) month is flited out when the monthly standardized anomaly of the IMI is greater (less) than 1.0 (−1.0). The standardized anomaly is computed by equation (1), and the selection results are shown in Table 1:

\[ X = \frac{(I - \bar{I})}{\sigma} \]  

where \( X \) is the standardized anomaly, \( I \) is the IMI of each month in JJAS (June-September) from 2000-2019, \( \bar{I} \) is the mean value of IMI, \( \sigma \) is the standard deviation of IMI.

<table>
<thead>
<tr>
<th>ISM strong months</th>
<th>ISM weak months</th>
</tr>
</thead>
<tbody>
<tr>
<td>2000.06</td>
<td>2002.07</td>
</tr>
<tr>
<td>2001.06</td>
<td>2009.06</td>
</tr>
<tr>
<td>2002.06</td>
<td>2009.08</td>
</tr>
<tr>
<td>2003.07</td>
<td>2012.06</td>
</tr>
<tr>
<td>2004.06</td>
<td>2014.06</td>
</tr>
<tr>
<td>2005.07</td>
<td>2014.08</td>
</tr>
<tr>
<td>2005.09</td>
<td>2015.07</td>
</tr>
<tr>
<td>2006.09</td>
<td>2015.08</td>
</tr>
<tr>
<td>2007.06</td>
<td>2015.09</td>
</tr>
<tr>
<td>2010.09</td>
<td>2018.09</td>
</tr>
<tr>
<td>2013.06</td>
<td>2019.07</td>
</tr>
</tbody>
</table>

Moisture availability and transport can be estimated and described by integrated water vapor transport (IVT) (Sousa et al. 2020). The IVT can be used to investigate the total amount of water vapor transport in an atmospheric column. The atmospheric data are used to calculate the IVT according to the equation (2):

\[ IVT(kg \cdot m^{-1} \cdot kg^{-1}) = \frac{1}{g} \int_{P_{sfc}}^{P_{top}} qV dp \]
where \(q\) is the specific humidity \((kg \cdot kg^{-1})\), \(V\) is the wind vector \((m \cdot s^{-1})\), \(g\) is the acceleration due to gravity \((m \cdot s^{-2})\). Previous studies usually set the top atmospheric pressure as 300 hPa (Chen et al. 2012; Feng and Zhou 2012; Sun et al. 2022), while the water vapor content above 300 hPa has shown a slight increasing trend in recent years due to the enhancement of evaporation over the TP (Xu et al. 2021; Zhang et al. 2016). Deep convection in the summer monsoon region can also transport low-level water vapor into the upper troposphere (Chen et al. 2012). Considering that there is still a slight but non-negligible amount of water vapor above 300 hPa, the top atmospheric pressure \(P_{\text{top}}\) in this study is set to be 100 hPa, and the surface pressure \(P_{\text{sfc}}\) is 1000 hPa.

Previous studies usually divided the TP boundary into several sub-boundaries, which can help to understand the detailed characteristics of the water vapor transport to the TP (Li et al. 2019; Li et al. 2022; Lin et al. 2018). According to the geometric shape of the TP, 10 boundaries are divided around the areas where the altitude is above 3000m, as shown in Fig.1. Since many studies have emphasized the importance of the southern boundary as a water vapor input boundary for the TP (Feng and Zhou 2012; Lin et al. 2016; Simmonds et al. 1999; Xu et al. 2014), this study focuses on boundaries 7, 8, 9, and 10 which are subdivided by the southern boundary as illustrated in Fig.1.

3 Results

3.1 Characteristics of water vapor transport and precipitation over the TP in ISM normal months

Figure 2 shows the IVT spatial distribution of diurnal variation over and around the TP in
The ISM normal months at (a) LT12, (b)LT13, (c)LT14, (d)LT15, (e)LT16, and (f)LT17. It is shown that the BoB located south of the TP is an essential external water vapor source for the TP. The warm-moist airflow originating from the BoB carries a large amount of water vapor to the TP across boundaries 8 and 7. Part of the airflow changes toward the west with a cyclonic rotation over northern India, reaching the TP across boundary 9. The mid-latitude westerlies are also water vapor transport paths that drive water vapor to the TP across boundary 10.

Figure 3 demonstrates the spatial distribution of hourly rain intensity (mm/hour\(^{-1}\)) over the TP and surrounding regions during ISM normal months at (a) LT12, (b)LT13, (c)LT14, (d)LT15, (e)LT16, and (f)LT17. An area with concentrated spatial precipitation in both ISM strong and weak months is regarded as a precipitation concentration region (PCR, red boxes in Fig. 3). It can be seen that precipitation mainly occurs in the southeastern of the TP (Sun et al. 2022), especially above the lakes like Siling Co and Zhari Namco. The PCR precipitation exhibits later diurnal peaks compared with the whole TP (Fig.6), as the former shows an afternoon peak with maximum precipitation at LT17 while the latter shows an afternoon peak at LT15 (Kukulies et al. 2020). Figure 5 demonstrates that precipitation in PCR has a good correlation with the water vapor transport across boundary 10, which is mainly affected by the mid-latitude westerlies. As is shown in Fig. 4b, the meridional water vapor fluxes across boundary 10 show a significant increasing trend from LT10 to LT18. Moreover, there is zonal water vapor transport through boundary ten from LT09 to LT17. Both of these conditions contribute to moisture supply for increased precipitation in PCR. For boundaries 8 and 9, mountains, e.g., the Himalayas, act as huge barriers to airflow. Driving by the sloping lateral surface heating over the southern of TP, the moisture of the airflow could turn into precipitation while climbing up the southern slope.
of the TP before reaching PCR. Contrary to the decreasing trend of precipitation in PCR during
the nighttime, the precipitation in front of the Himalayas shows a significant ascending trend
during the nocturnal hours (LT00-04), which has been reported by previous studies (Barros et
of Barros and Lang (2003), the atmospheric instability and moisture over the Himalayas
increase gradually during the day and are then released around midnight, leading to the
nocturnal peak in rainfall along the Himalayas.

3.2 Difference of IVT and precipitation over the TP between ISM strong months
and ISM normal months

Figure 8 shows the spatial distribution differences for precipitation and IVT between the
ISM strong and ISM normal months over the TP and surroundings at (a) LT08, (b)LT09,
(c)LT10, (d)LT17, (e)LT18, and (f)LT19. When the ISM is enhanced, precipitation increases
slightly in most parts of the TP. The diurnal cycle of the precipitation differences between the
ISM strong and normal months in PCR is displayed in Fig. 9. For the PCR, the ISM strong
months precipitation is greater than ISM normal months in each hour of the day. The diurnal
cycle is characterized by a morning peak at LT09 and an early evening peak at LT18. Figure 10
depicts the spatial distributions of correlation coefficients of precipitation difference in PCR
with IVT difference over the TP and the surroundings between ISM strong and normal months.
The water vapor transported across boundaries 8, 9, and 10 shows positive correlations with the
PCR precipitation. Since water vapor outflows from the TP to the south across these four
boundaries from LT00 to LT09 (Fig. 11b), the peak of increased precipitation in PCR at LT09
is mainly caused by the meridional water vapor input. As shown in Fig. 11a, the water vapor can be meridionally transported into the TP through boundary 10 from LT00 to LT09. At the same time, the meridional water vapor fluxes across boundary 9 also show a significant increasing tendency from LT04 to LT09. Figures 7a, 7b, and 7c illustrate that the enhanced northwest airflows can transport the water vapor into the TP. Both of them contribute to the increase of the PCR precipitation from LT00 to LT09. Then in the afternoon (LT15 – LT18), the PCR precipitation increases again and reaches its daily maximum at LT18. On the one hand, the west-east component of water vapor flux across boundary 9 increases during the same time (Fig. 11a). Meanwhile, it is noticed that the IVT around boundary 7 shows a slightly enhanced trend during this time. The enhanced southwest air flows help the water vapor into the TP, so the meridional water vapor fluxes across boundary 7 show an increasing tendency from LT15 to LT20. While considering that the enhanced westerly airflows dominate the TP, the moisture transported across boundary 7 may be challenging to reach PCR. Therefore, water vapor transport across boundary 7 may contribute little to the increased precipitation in PCR (Fig. 7d). On the other hand, the zonal water vapor flux across boundary 8 shows a notable increasing trend during this time (Fig. 11b). As shown in Figs. 7d and 7e, the IVT increased significantly near boundary 8, which is beneficial to the water vapor transport from the northern BOB to the TP, and then the water vapor can be carried to PCR by the airflow.

3.3 Difference of IVT and precipitation between ISM weak months and ISM normal months

Figure 13 shows the spatial distribution of precipitation differences and IVT differences
between the ISM weak and normal months at (a) LT06, (b)LT07, (c)LT08, (d)LT20, (e)LT21, and (f)LT22. It shows that the TP precipitation changes in the ISM weak months are smaller than in the ISM strong months. It can be seen from Fig. 14 that the PCR precipitation in ISM weak months is smaller than that in ISM normal months for most of the day. The precipitation in PCR only increases in the evening (LT19-22) and reaches a peak at LT21. Figure 15 shows the distribution of correlation of differences of the PCR precipitation and the IVT over the TP and the surroundings between ISM weak and normal months. As shown in Fig. 15, the water vapor transport across boundaries 10 and 8 correlates well with the PCR precipitation. First, the IVT increases around boundary 10 (Fig. 7), suggesting that water vapor can be transported from the Arabian Sea to the TP across boundary 10. It can be seen in Fig. 16b that the zonal water vapor transports out across boundaries 7, 8, and 9 except boundary 10 during the whole day. Both meridional and zonal water vapor flux across boundary 10 show increasing tendencies from afternoon to evening (LT16-22), contributing to the PCR precipitation increase. Secondly, there is a water vapor transport channel in boundary 9, around 29° N, 82° E. From LT14 to LT18 (Fig. 12), the increased IVT implies that the water vapor transport between the TP and the Arabian Sea is active, resulting in the increase of meridional water vapor transport across boundary 9, which also contributes to the precipitation increase in PCR. Finally, the lower terrain near boundary 8 provides a favorable condition for the water vapor entrance of the TP. Since the IVT around boundary 8 is enhanced (Fig. 12), boundary 8 is an efficient boundary for the water vapor transport between the TP and the south during the day. The meridional water vapor transport across boundary 8 shows an increasing trend during LT14-LT19, contributing moisture to the maximum precipitation in PCR at LT21.
4 Discussion and Conclusions

This study focuses on the effects of the ISM on the characteristics of water vapor transport across the southern boundary of the TP and its impacts on the change of the precipitation over the TP. The ERA5 reanalysis data from 2000 to 2019 and CMORPH precipitation products are used to investigate this issue. The main conclusions are summarized as follows.

The water transport of the TP southern boundary is affected by both mid-latitude westerlies and ISM. The precipitation in PCR shows a positive correlation relationship with the water vapor transport across boundary 10, which is influenced by mid-latitude westerlies, no matter whether the ISM is strong or weak. In ISM weak months, the correlation between the PCR precipitation and the water vapor transport across boundaries 9 and 8 originating from the BoB and the Arabian Sea is weaker compared to ISM strong months. The ISM-dominated boundary 7 is a water vapor source boundary in both ISM strong and weak months. PCR is located northwest of boundary 7. The circulation pattern over the TP is dominated by the westerly winds in ISM strong and weak months. As a result, the southwesterly warm-moist airflows flowing across boundary 7 are challenged to reach the PCR.

Compared with the precipitation of the ISM normal months, the PCR precipitation in the ISM strong months is greater in each hour of the day and reaches its peak at LT18. The enhanced IVT around boundaries 8 and 9 indicates that the airflow carries moisture from the BoB to the plateau and contributes to the increased PCR precipitation. Meanwhile, the zonal water vapor transport across boundary 10 also brings moisture to the PCR and is beneficial to increasing the PCR precipitation.

For the ISM weak months, the precipitation is smaller than that in ISM normal months for
most of the day. The PCR precipitation is mainly affected by water vapor transport across boundary 10, primarily driven by mid-latitude westerlies. Part of water vapor can also be transported meridionally to PCR via boundary 8. Although both mid-latitude westerlies and ISM can carry water vapor to the TP direction in ISM strong months, the water vapor that is efficiently transported into the TP is less in ISM strong months than in ISM weak months, which agrees with the study of Sugimoto et al. (2008). When the ISM is strong, there is a cyclonic circulation with a low-pressure area over northern India. It will change the moisture transportation direction from the southwest to the southeast. The southeasterly air flows converge with the enhanced northwesterly flows in the north of India, causing a large amount of precipitation. A lot of water vapor could be consumed by these precipitation events before it reaches the TP. For the ISM weak months, water vapor can be transported by the enhanced northwest flows from the Arabian Sea and reach the TP more efficiently. Therefore, a large amount of water vapor can be transported to the TP in ISM weak months.

It is noticed that less water vapor is transported to PCR, but the precipitation is greater in ISM strong months. One possible reason for this is that PCR locates in the mid-western region of TP, where the precipitation recycling ratio (PRR) is relatively high. The PRR is the local evaporation that contributes to the local precipitation (Dirmeyer and Brubaker 2007). The higher PRR value represents a greater contribution of local moisture to local precipitation. Since precipitation in PCR is greater than the TP in other regions in both the ISM strong and weak months, precipitation in PCR may not depend much on external water vapor transport. Zhang et al. (2017) also found that the estimated PRR increased at a rate of 1.4% decade$^{-1}$ with the
precipitation increased in the mid-western TP.

In the procedure of the study, some issues still deserve further analysis. For example, quantify the water vapor entered from the southern boundary and transported to the PCR, figure out the factors that influence PCR precipitation besides external water vapor transport. In future work, more detailed cloud characteristics (e.g., cloud macro- and micro-physical features) can be investigated to understand the mechanism of the ISM affecting the TP cloud and precipitation processes.

References


0493(2003)131<1408:MTMITH>2.0.CO;2

1

Chen B, Xu X De, Yang S, Zhang W (2012a) On the origin and destination of atmospheric moisture

https://doi.org/10.1007/s00704-012-0641-y


https://doi.org/10.5194/acp-12-5827-2012


https://doi.org/10.1175/JHM557.1


https://doi.org/10.1038/ncomms10925


https://doi.org/10.1007/s00704-015-1645-1


https://doi.org/10.1016/j.atmosres.2021.105574


https://doi.org/10.1029/2019JD031297


https://doi.org/10.1175/JCLI-D-19-0348.1


https://doi.org/10.2151/jmsj.86.935


https://doi.org/10.1016/j.scitotenv.2022.153545


https://doi.org/10.1175/1520-0442(2001)014<4073:IVOTAS>2.0.CO;2


https://doi.org/10.1002/joc.4609


https://doi.org/10.1002/2016JD025515


https://doi.org/10.1002/rog.20023


https://doi.org/10.1038/nclimate1580


https://doi.org/10.1175/JHM-D-18-0094.1


**Statements & Declarations**

**Funding**

This study was supported under the National Science Foundation of China (Grant Nos. 42075067, 42075068), and the Second Tibetan Plateau Scientific Expedition and Research (STEP) program (Grant No. 2019QZKK0105).

**Competing Interests**

The authors have no relevant financial or non-financial interests to disclose.

**Author Contributions**

The authors confirm contribution to the paper as follows: study conception and design: Jinghua Chen and Tianyu Liu; data collection: Tianyu Liu and Kai Yang; analysis and interpretation of
results: Tianyu Liu, Jinghua Chen and Liping Deng; draft manuscript preparation: Tianyu Liu and Jinghua Chen. All authors reviewed the results and approved the final version of the manuscript.

All authors reviewed the manuscript.

**Data Availability**

The ERA5 reanalysis data analyzed during the current study are available in the Copernicus Climate Data Store (https://climate.copernicus.eu/climate-reanalysis) and the hourly CMORPH precipitation product analyzed during the current study are available in National Meteorological Information Center (http://data.cma.cn/). The IMI analyzed during the current study are available in Monsoon Monitoring Page (http://apdrc.soest.hawaii.edu/projects/monsoon/seasonal-monidx.html).

**Figures**

![Figure 1](image)

**Fig. 1** The area boundaries of the TP. The shaded color denotes the altitude(m), and the red lines denote the study boundary for the vapor transport between the TP and its surrounding areas. The boundaries are marked as boundary 1 (26.5°N-40.5°N), boundary 2 (73°E-81°E, 36.5°N-40.5°N),
boundary 3 (81°E-98°E, 36.5°N-40°N), boundary 4 (98°E-103.5°E, 37°N-40°N), boundary 5 (103.5°E-105°E, 34°N-37°N), boundary 6 (100°E-105°E, 26.5°N-34°N), boundary 7 (95°E-100°E, 26.5°N-28.5°N), boundary 8 (89°E-95°E, 26.5°N-28.5°N), boundary 9 (80°E-89°E, 26.5°N-29°N), boundary 10 (70°E-80°E, 29°N-37°N)

Fig. 2 IVT (units: kg·m⁻¹·s⁻¹) spatial distribution of diurnal variation over and around the TP in the ISM normal months at (a) LT14, (b)LT15, (c)LT16, (d)LT17 (e)LT18, and (f)LT19. The bold black lines denote the topographic height of 3000 m.
Fig. 3 Spatial distribution of hourly rain intensity (contour; units: mm·h⁻¹) and IVT (vectors; units: m·s⁻¹) over the TP and surrounding regions during ISM normal months at (a) LT12, (b) LT13, (c) LT14, (d) LT15, (e) LT16, and (f) LT17. The bold black lines denote the topographic height of 3000 m. The PCR (30-32.5° N, 85-90° E) location is shown in red boxes.
**Fig. 4** Diurnal cycle of (a) meridional and (b) zonal water vapor flux (units: 10^{-4} \text{ g·m}^{-1}·\text{Pa}^{-1}·\text{s}^{-1}) transported across boundaries 7, 8, 9, 10 during ISM normal months.

**Fig. 5** Distribution of correlation between precipitation in PCR and IVT over the TP and the surrounding regions during ISM strong months (coloring represents the correlation coefficient, where the correlation coefficient greater than 0.150 is considered to be more than 90%). The bold black lines denote the topographic height of 3000 m. The black box indicates the location of PCR (30-32.5° N, 85-90° E).
Fig. 6 Diurnal cycle of the precipitation (units: mm·h⁻¹) in PCR during ISM normal months

Fig. 7 The IVT (units: kg·m⁻¹·s⁻¹) spatial distribution of diurnal variation of the difference between the ISM strong months and ISM normal months at (a) LT08, (b) LT09, (c) LT10, (d) LT17, (e) LT18, and (f) LT19. The bold black lines denote the topographic height of 3000 m
Fig. 8 Spatial distribution of precipitation differences (contour; units: mm·h$^{-1}$) and IVT differences (vectors; units: kg·m$^{-1}$·s$^{-1}$) between the ISM strong and ISM normal months over the TP and surroundings at (a) LT08, (b) LT09, (c) LT10, (d) LT17, (e) LT18, and (f) LT19. The bold black lines denote the topographic height of 3000 m. The PCR (30-32.5°N, 85-90°E) location is shown in red boxes.
**Fig. 9** Diurnal cycle of the precipitation (units: mm·h⁻¹) difference between the ISM strong months and normal months in PCR.

**Fig. 10** Spatial distributions of correlation coefficients of precipitation difference in PCR with IVT difference over the TP and the surroundings between ISM strong months and ISM normal months. The bold black lines denote the topographic height of 3000 m. The black box indicates the location of PCR (30-32.5°N, 85-90°E).
Fig. 11 Diurnal cycle of the differences of (a) meridional and (b) zonal water vapor flux (units: $10^{-4} \text{g} \cdot \text{m}^{-1} \cdot \text{Pa}^{-1} \cdot \text{s}^{-1}$) transport across boundaries 7,8,9,10 between ISM strong months and ISM normal months.

Fig. 12 The IVT (units: kg $\cdot$ m$^{-1} \cdot$ s$^{-1}$) spatial distribution of diurnal variation of the difference between
the ISM weak months and ISM normal months at (a) LT06, (b) LT07, (c) LT08, (d) LT20, (e) LT21, and (f) LT22. The bold black lines denote the topographic height of 3000 m.

**Fig. 13** Spatial distribution of precipitation differences (contour; units: mm·h\(^{-1}\)) and IVT differences (vectors; units: kg·m\(^{-1}\)·s\(^{-1}\)) between the ISM weak and ISM normal months over the TP and surroundings at (a) LT06, (b) LT07, (c) LT08, (d) LT20, (e) LT21, and (f) LT22. The bold black lines denote the topographic height of 3000 m. The PCR (30-32.5°N, 85-90°E) location is shown in red boxes.
Fig. 14 Diurnal cycle of the precipitation differences (units: mm·h⁻¹) between the ISM weak months and normal months in PCR.

Fig. 15 Spatial distributions of correlation coefficients of precipitation difference in PCR with IVT difference over the TP and the surroundings between ISM weak months and ISM normal months. The bold black lines denote the topographic height of 3000 m. The black box indicates the location of PCR (30-32.5°N, 85-90°E)
**Fig. 16** Diurnal cycle of the differences of (a) meridional and (b) zonal water vapor flux (units: $10^{-580} \cdot \text{g} \cdot \text{m}^{-1} \cdot \text{Pa}^{-1} \cdot \text{s}^{-1}$) transport across boundaries 7,8,9,10 between ISM weak months and ISM normal months.