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Influence of atmospheric heat sources over the Tibetan Plateau and the tropical western North Pacific on the inter-decadal variations of the stratosphere-troposphere exchange of water vapor

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This study investigates the Stratosphere-Troposphere Exchange (STE) of water vapor, emphasizes its inter-decadal variations over Asia in boreal summer, and discusses the influences of atmospheric heat sources over the Tibetan Plateau and the tropical western North Pacific (WNP) on them by using the Wei method with reanalysis data from the European Centre for Medium-Range Weather Forecasts (ECMWF) for the years of 1958—2001. The climatology shows that the upward transport of water vapor across the tropopause in boreal summer is the most robust over the joining area of the South Asian Peninsula and Indian-Pacific Oceans (defined as AIPO). The upward transport over there can persistently convey the abundant water vapor into the stratosphere and then influence the distribution and variation of the stratospheric water vapor. The analysis shows that inter-decadal variations of the water vapor exchange over the AIPO are significant, and its abrupt change occurred in the mid-1970s and the early 1990s. In these three periods, as important channels of the water vapor exchange, the effect of Bay of Bengal-East Asia as well as South China Sea was gradually weakening, while the role of the WNP becomes more and more important. Further studies show that atmospheric heat sources over the Tibetan Plateau and the WNP are two main factors in determining the inter-decadal variations of water vapor exchange. The thermal influences over the Tibetan Plateau and the WNP have been greatly adjusted over the pass 44 years. Their synthesis influences the inter-decadal variations of the water vapor exchange by changing the Asian summer monsoon, but their roles vary with time and regions. Especially after 1992, the influence of heat source over the Tibetan Plateau remarkably weakens, while the heat source over the WNP dominates the across-tropopause water vapor exchange. Results have important implications for understanding the transport of other components in the atmosphere and estimating the impact of human activities (emission) on global climate.

Tibetan Plateau, tropical western North Pacific, atmospheric heat source, STE, inter-decadal variations

Water vapor in the Upper Troposphere and Lower Stratosphere (UTLS) is a key greenhouse gas which exerts a major influence on the energy balance of the earth-atmosphere system. Water vapor is also important for atmospheric chemistry as the primary source of the hydroxyl radical, which changes ozone concentration and eliminates the contamination in the atmosphere by participating in the photochemical processes in UTLS. The water vapor balance in the UTLS is dominated by many factors among which the Stratosphere-Troposphere Exchange (STE) of water vapor is the decisive

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one. For example, the stratospheric dryness results from dehydration of air across the cold tropical tropopause[1]. Detailed process studies of the water vapor transport may suggest the transport of other components in the atmosphere, and might help to estimate the impact of human activities (emission) on global climate. In addition, mass exchange across the tropopause is a key process in the atmospheric model, but up to now, this process can not be completely understood and modeled. Consequently, it is of critical importance to study the STE of water vapor.

The STE of water vapor has received attention over the pass half a century. In order to explain the phenomenon of stratospheric dryness, Brewer[1] first concluded that all air entering the stratosphere must be freeze-dried near the tropical tropopause. This overall view of stratospheric circulation has been confirmed by the subsequent observations and researches[2]. However, there exists a quantitative mismatch. This prompted Newell and Gould-Stewart[3] to propose the hypothesis that air enters the stratosphere only in restricted areas and at certain time of the year. These regions, termed "stratospheric fountains", were said to be located over the Indonesian region between November and March and over the Bay of Bengal (BOB) during boreal summer. Although this hypothesis provides an effective explanation for understanding the stratospheric dryness, it has been more and more questioned by the wider application of satellite data and the in-depth study on the STE[2,4]. Rosenlof[4] pointed out that other processes must also be important.

In 1995, Chen[5] first emphasized the importance of the Asian summer monsoon circulations to influence the STE. The subsequent studies have shown that the Asian monsoon region is a main channel through which water vapor and other constituents in the troposphere can enter the lowermost stratosphere and have strong impact on the stratospheric chemistry and radiation budget[6–13]. Hence, identifying and investigating the dynamical and physical processes of the water vapor transport, especially its long-term processes, are critical for understanding the radiative force affecting the global climate change and the stratospheric chemistry and hydration. The above conclusions, however, were drawn only from the short-term data. The quantitative process of the water vapor transport is a major issue that remains unsolved, especially its decadal characteristics and effect mechanism need to be further studied.

As early as the 1950s, Tao et al.[14] showed that monsoon is a synthetical result of several factors such as radiative change, adjustment of atmospheric general circulation, land-sea distribution and topography. Further, Zeng and Li[15] pointed out that the tropical monsoon is driven by the planetary thermal convection and the differences between the thermal characters of the earth surface. Since the monsoon which has important impact on the STE of water vapor is closely related to heat source and the land-sea thermal contrast, it is natural that the water vapor transport is in relation to heat source and the land-sea thermal contrast. It is well recognized that the Tibetan Plateau is the outstanding heat source in boreal summer, and its thermal variations play an important role in affecting the Asian monsoon circulation, the formation of ENSO, and the transports of water vapor and mass[7,11,16–20]. On the other hand, the Rossby wave triggered by the thermal convection over the tropical western North Pacific (WNP) contributes to the atmospheric general circulation over Asia[21–24]. Furthermore, the heat sources over the above two regions were characterized by interdecadal variations in the past half a century. The following questions have been naturally raised: Have the interdecadal variations of the heat source an important impact on the STE of water vapor? If so, what is the effect mechanism? Up to now, there has been little published work about the long-term variations of the water vapor transport between the troposphere and the stratosphere, and its relationships with the atmospheric heat sources are poorly understood. This study will, therefore, address the STE of water vapor with a focus on its inter-decadal variations, and discuss the physical mechanism of inter-decadal variations from the impacts of heat sources over the Tibetan Plateau and the tropical WNP. Results will be important for understanding the STE.

1 Data and methodology

1.1 Data

The data used in this study include the following sources: (1) 6-hour reanalysis data provided by European Centre for Medium-Range Weather Forecasts (ECMWF) for the years of 1958－2001 (ERA40). This dataset is used to drive the Wei method in this study; (2) Monthly mean ERA40. This dataset is used to analyze the variations of atmospheric circulation.

All data were stratified into four seasons: December-

February (DJF), March-May (MAM), June-August (JJA), and September-November (SON). Although all seasons were analyzed at the beginning, our discussion will focus primarily on the JJA season, which shows the strongest evidence for the water vapor transport from the troposphere to the stratosphere.

1.2 Methodology

(1) Mathematical formulation of STE. In 1987, Wei[25] presented the Eulerian diagnostic model of the STE. Let $F(f)$ represent the cross-tropopause transport term:

$$F(f)=\left[\rho J_{\eta} f\left(\frac{d\eta}{dt}-\frac{\partial\eta}{\partial t_{z_0}}-U\cdot\nabla z_0\eta\right)\right]_{\eta B}, \qquad (1)$$

where $\eta=\eta(z)$ is a generalized vertical coordinate with $J=\partial z/\partial\eta$, f is a specific physical property, $\frac{\partial\eta_B}{\partial t_\eta}=\frac{\partial\eta}{\partial t_{\eta B}}$ and the subscripts η_B and Z_0 are interchangeable, U is the horizontal wind vector, and ρ is the density. The first term on the right-hand side of eq. (1) represents the exchange between the stratosphere and troposphere due to the vertical velocity at tropopause; the second term results from the temporal movement of the tropopause in the η-space; the last term is due to the variation of η along the tropopause surface. This model is independent of the coordinate, and can be used to calculate the flux of optional mass and variables (such as water vapor, ozone) cross the optional parameterized surface.

One can obtain different versions of Wei's formula depending on the choice of the vertical coordinate. If choose the isobaric coordinates and let $f = 1$, the cross-tropopause mass transport would be

$$F(m)=\frac{1}{g}\left(-\omega+V_h.\nabla P_{tp}+\frac{\partial P_{tp}}{\partial t}\right)$$
$$=\left(-\frac{\omega}{g}+\frac{1}{g}V_h.\nabla P_{tp}\right)+\frac{1}{g}\frac{\partial P_{tp}}{\partial t}=F_{AM}+F_{TM}, \qquad (2)$$

and if let f=q, the cross-tropopause water vapor transport $F(q)$ is expressed as

$$F(q)=\frac{q}{g}\left(-\omega+V_h.\nabla P_{tp}+\frac{\partial P_{tp}}{\partial t}\right)$$
$$=\left(-\frac{q\omega}{g}+\frac{q}{g}V_h.\nabla P_{tp}\right)+\frac{q}{g}\frac{\partial P_{tp}}{\partial t}=F_{AQ}+F_{TQ}, \qquad (3)$$

where ω is the pressure vertical velocity, V_h is the horizontal wind vector, P_{tp} is the pressure of the tropopause, q is the specific humidity, and g is the acceleration due to gravity. More detailed presentations of the manipulations involved can be found in ref. [25]. $F_{AM}(F_{AQ})$ represents the air-mass (water vapor) exchange due to the horizontal and vertical air motion. $F_{TM}(F_{TQ})$ arises from the tropopause motion. Positive values of $F(m)$ and $F(q)$ imply transport from the troposphere to the stratosphere; negative values imply transport in the opposite direction. We have introduced the numerical modifications (referred to as the "advection method") developed by Siegmund et al.[26] into eqs. (2) and (3) to reduce errors associated with the practical problems of time and space differencing. Siegmund et al. fully discussed the mathematics and the reduction of the resulting spurious noise. The above diagnostic model has been used widely to evaluate the flux of air cross the tropopause[8,26−31]. However, it should be noted that the resolution may bring an additional uncertainty into the analysis since there exists the smaller-scale exchange between troposphere and stratosphere. The previous studies using ECMWF[26] and NCEP/NCAR data[8] have shown that the coarse resolution will obtain the falsely results which are larger than the actual ones. The diagnosed instantaneous STE can not notably change if the temporal resolution of the circulation data is decreased from 3h to 6h. A further resolution decreases to 12h, however, generally leads to quite different results for the instantaneous STE. The horizontal resolution increase from 8°×10° to 4°×5° resulted in almost a doubling of the evaluated net transport. A further increase to 2.5°×2.5°, on the other hand, changed the results by only about 10%. Hence, it suggests that 6-hourly data available on a 2.5°×2.5° horizontal grid are appropriate. In this study, 6-hourly ERA40 data available on a 2.5°×2.5° horizontal grid are used to drive the Wei method, and then proceeded into the corresponding resolution.

(2) The choice of the tropopause. There are many methods of tropopause definition with the man-made factors. Various tropopause definitions have been used according to the different research purpose, including the conventional thermal definition based on lapse-rate criteria, a dynamical definition based on a potential-vorticity (PV) threshold, and a chemical definition based on ozone gradient and values. Considering that the re-

gion referred to in this study is the strongest region of heat source in summer and has an especial thermal condition, the thermal definition of the tropopause is used for evaluating STE here. Following the World Meteorological Organization[32], the thermal method defines the tropopause as the lowest layer at which the temperature lapse rate is less than 2℃·km^{-1} for a depth of at least 2 km. In fact, the STE of water vapor in dynamical tropopause is also examined. The results showed that the exchanges in thermal and dynamical definitions are similar to each other, and the first definition can not only keep the continuity of the tropopause but also avoid the uncertainty from the different thresholds in the dynamical definition.

(3) Mathematical formulation of heat source. The atmospheric heat can not be obtained by direct observation, so two theories are commonly used to estimate it. One is to regard the net heat source as remanding term in thermodynamic conservational equation according to energy balance, and the other is to calculate every component of thermodynamics, respectively. This study will adopt the first theory (reverse method) to estimate the atmospheric heat source.

As shown by Yanai[33] and Ding[34], the apparent heat source Q1 is computed from

$$Q_1 = c_p\left[\frac{\partial T}{\partial t} + V\cdot\nabla T + \left(\frac{P}{P_0}\right)^{\kappa}\omega\frac{\partial\theta}{\partial p}\right], \qquad (4)$$

where P_0=1000 hPa, κ=R/C_P, R and C_P are, respectively, the gas constant and the specific heat at constant pressure of dry air, and θ is the potential temperature.

Integrating eq. (4) from 100 hPa to the surface pressure P_S, we obtain

$$\langle Q_1\rangle = \frac{1}{g}\int_{100}^{p_s} Q_1 \mathrm{d}p = LP + Q_s + \langle Q_R\rangle, \qquad (5)$$

where P, Q_S, and $\langle Q_R\rangle$ are the precipitation rate, the sensible heat flux, and the radiative heating rate, respectively.

In addition, we investigate the inter-decadal variations of the water vapor transport by empirical orthogonal function (EOF) analysis, moving t-test (MTT) method, and composite analysis. We then remove the interannual component by 9-point Guassian-type filter to study the relationship between the water vapor exchange and the atmospheric heat sources on inter-decadal time scale. Finally, we discuss the possible mechanism for influence of the atmospheric heat sources on the STE using regression analysis.

2 44-year climatology of the STE of water vapor in the Northern Hemisphere

The climatology of the STE of water vapor is investigated first. Seasonal variations of the water vapor transport in the Northern Hemisphere are shown in Figure 1. It is clear that the distribution of water vapor transport between the troposphere and the stratosphere basically corresponds to thermal-driven Brewer-Dobson circulation model. In the tropics, air ascends from the troposphere to the lower stratosphere, so here the STE is mainly characterized by the transport from the troposphere to the stratosphere, which plays an important role in distribution and variation of the stratospheric water vapor. In the extra-tropics, air descends from the stratosphere to the troposphere with a result that the STE is mainly from the stratosphere to the troposphere. The extra-tropical STE has important influences not only on distribution and variation of the upper-tropospheric water vapor but also on the triggering of strong convection and precipitation by the formation of geopotential instability with the lower-level wet and warm flows. It should be noticed that in the breaks between the polar tropopause and the tropical tropopause, which is indicated by contours of tropopause pressure = 150 hPa in Figure 1 and consistent with the results by MAXOBEP[35], there exists the strongest water vapor transport from the stratosphere to the troposphere the whole year round. Especially in winter and summer seasons, the position of tropopause break basically corresponds to the extreme axis of stratosphere-to-troposphere water vapor transport. It indicates that the tropopause break plays an important role in the water vapor transport from the stratosphere to the troposphere. However, since Gettelman et al.[30] pointed out that the Wei diagnostic model is particularly sensitive to input data errors near the jet stream (its position in the subtropics corresponds in the rough to that of the tropopause breaks), the authenticity of STE near the tropopause breaks needs to be further studied.

In addition, the STE of water vapor also exhibits the strong zonal asymmetry and the regionality due to the influences from the atmospheric circulation and other outer forces. A first look is put in the tropics. Figure 1(a)

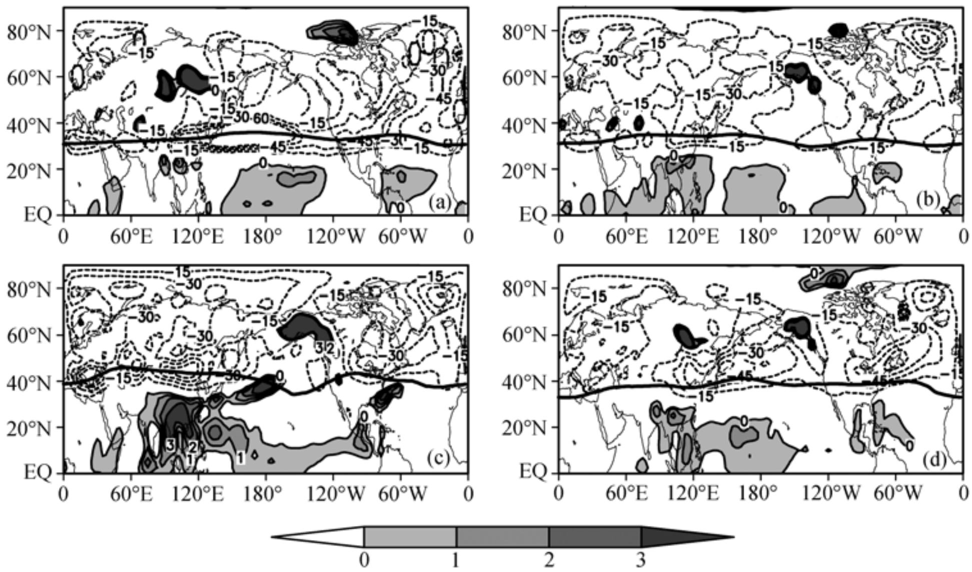

Figure 1 Seasonal distributions of STE of water vapor in the Northern Hemisphere during 1958—2001 (unit: 10^{-9} kg·m^{-2}·s^{-1}). (a) Winter (DJF); (b) Spring (MAM); (c) Summer (JJA); (d) Autumn (SON). Thick solid lines show contours of tropopause pressure = 150 hPa.

shows that the positive maxima of water vapor transport are located over the central Pacific and the western Atlantic in winter. As season changes, the positive areas gradually extend northwest and the positive center moves over the Asian continent. In summer (Figure 1(c)), the joining area of the South Asian Peninsula and Indian-Pacific Oceans becomes the strongest transport region from the troposphere to the stratosphere in the Northern Hemisphere. This region seems similar to the joining area of Asia and Indian-Pacific Oceans (AIPO) presented by Wu et al[36], and herein is also defined as AIPO. In fact, the water vapor transport over this region is also the strongest in the globe (figure omitted). The AIPO are consistent with the location of channel of mass transport found by Cong et al.[7] and Guo et al.[13], and two positive centers over the northern BOB and the southeastern Tibetan Plateau correspond to the results by Cong et al.[7] based on 19-year NCEP data. It suggests that from the long-term view, the AIPO are foremost in mass/water vapor transport from the troposphere to the stratosphere, which may be related to the summer monsoon prevailing over the region. Up to autumn (Figure 1(d)), the positive maximum over Asia rapidly weakens and southward withdraws, so that the positive coverage over Asia greatly reduces. In extra-tropics, the STE is mainly characterized by the stratosphere-to-troposphere transport but there still exists the outstanding upward transport across tropopause over one or two regions. However, it will not be further studied in this study.

The above results show that the upward transport of water vapor across the tropopause in boreal summer is robust over the joining area of AIPO. The upward transport over there can persistently convey the abundant water vapor to the stratosphere and then influence the distribution and variation of the stratospheric water vapor. At the same time, such transport will also convey the tropospheric contamination to the stratosphere. In this sense, the AIPO plays an active role in the stratospheric radiation, photochemistry and global climate. This study will therefore focus on the STE of water vapor in boreal summer, and discuss its characteristics over the AIPO (40°—180°E, 0°—40°N).

3 Inter-decadal variations of the STE of water vapor

In order to investigate the STE of water vapor over the AIPO in boreal summer, EOF analysis is performed for the 44-year field of water vapor transport. The first three EOFs and corresponding time series, shown in Figure 2, explain about 12%, 9%, and 7% of the total variance. Together, the leading three EOFs account for nearly one third changes of water vapor transport in boreal summer.

Figure 2(a) shows that the EOF1 has a zonal negative-positive-negative structure in its spatial distribution, that is to say, the cross-tropopause flux anomalies of water vapor over 40°—80°E and 120°—180°E are out of phase with those over 80°—120°E. Time series cor-

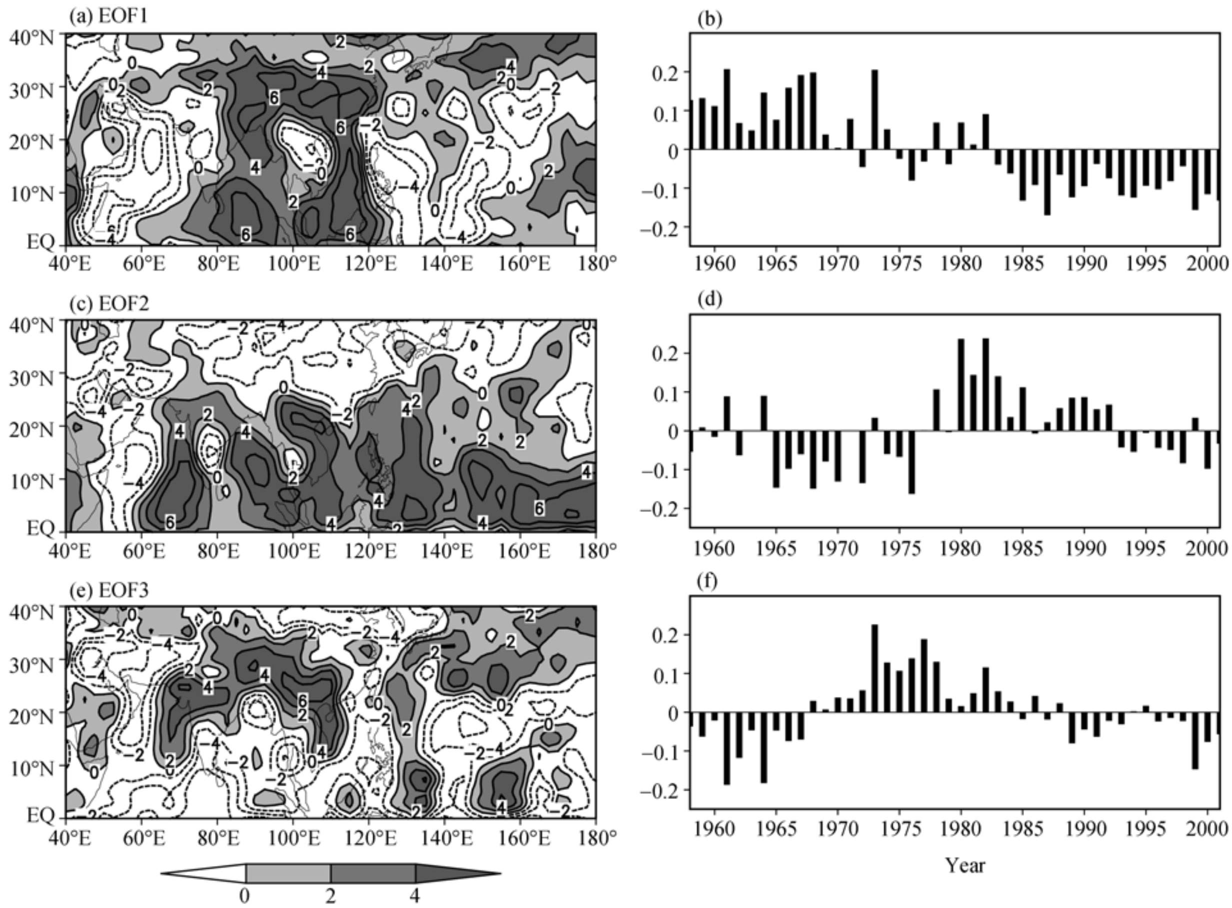

Figure 2 The first (a), second (b) and third (c) leading EOF (empirical orthogonal function) modes, and the corresponding time series ((b), (d), (f)) of the STE anomalies of water vapor over Asia in boreal summer during 1958—2001. Vertical axis is time coefficient for (b), (d) and (f).

responding to the first EOF show that the cross-tropopause fluxes of water vapor are characterized by gradually weakening, with positive anomalies before 1974, negative ones after 1983, and interlaced ones between 1975 and 1982. The EOF2 is dominated by the out-phase structure in north-south direction with a boundary at about 25°N. Time series for this mode remarkably differ from those of the first mode, with negative anomalies before the mid-1970s and after the early 1990s, and positive ones between them. To test the abrupt point of time series corresponding to the second EOF, MTT method is further adopted here. Figure 3 shows that the water vapor transport has mainly two below-normal epochs, one between 1965 and 1977 and the other between 1993 and 2001. The water vapor transport also has an above-normal epoch during 1978—1992. It should be indicated that the abrupt point in 1964 is not evident and the water vapor transport in that year nearly equals to the normal value. The third EOF and its time series (Figure 2(e)—(f)) exhibit an interdecadal

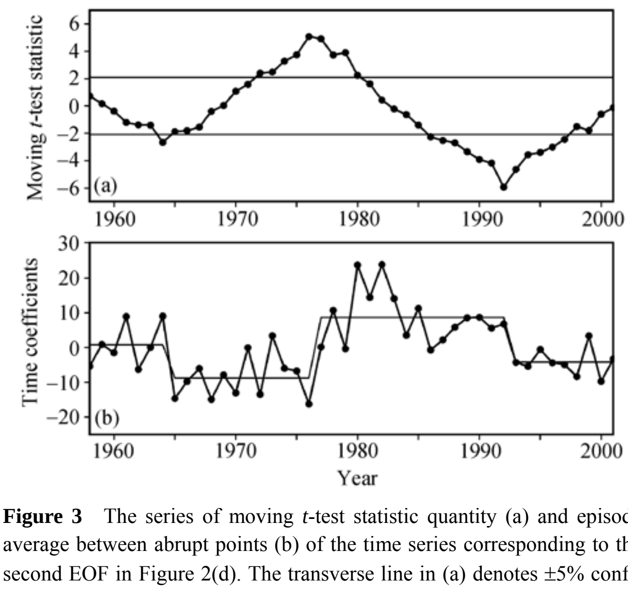

Figure 3 The series of moving *t*-test statistic quantity (a) and episode average between abrupt points (b) of the time series corresponding to the second EOF in Figure 2(d). The transverse line in (a) denotes ±5% confidence level.

nature with positive anomalies over the Marwar Plateau-Tibetan Plateau-Yunan-Guizhou Plateau region and the West northern Pacific and with negative anomalies over the other regions. The abrupt points of this mode occur in late 1960s and in late 1980s, respectively. According to North criterion[37], the third mode has no physical significance, so the selection of the stable periods only depends on the first two modes. Considering the following two facts that time series for the EOF1 have such a trend as gradually weakening and those for the EOF2 are characterized by interdecadal variability, we divided the data into the following three stable periods: 1958—1977, 1978—1992, and 1993—2001.

We now examine the interdecadal variations of the STE of water vapor over the summer Asia. Figure 4 shows the cross-tropopause flux anomalies of water vapor over the AIPO in boreal summer during the periods of 1958—1977, 1978—1992, and 1993—2001. Similar to the EOF1, the water vapor transport over the summer Asia during 1958—1977 (Figure 4(a)) is dominated by a zonal negative-positive-negative pattern, that is to say, negative anomalies are located over Arabian Sea-India, northern Indo-China Peninsula, the Philippines and the WNP east to the Philippines, while positive anomalies are over the Arabian Peninsula, BOB, Chinese mainland-South China Sea-southern Indo-China Peninsula, and the northern WNP. In this period, the cross-tropopause transports over the BOB and East Asian monsoon region are stronger than normal. In comparison with the first period, the cross-tropopause flux anomalies turn to opposite sign over most of the AIPO in the later period (Figure 4(b)) except for the positive anomalies over northern BOB and South China Sea-southern Indo-China Peninsula. Combining the difference between 1978—1992 and 1958—1977 shown in Figure 4(d), it is clear that the interdecadal change over the most regions is significant in these two periods. During the period of 1993—2001 (Figure 4(c)), the positive anomalies over northern BOB and South China Sea-southern Indo-China Peninsula vanish, so that the negative anomalies with greater amplitude than the former period prevail in the region of 80°—120°E and 0°—35°N and most of them surpass the significant at 95% confidence level (Figure 4(e)). The positive anomalies over the Arabian Sea eastward extend and cause the

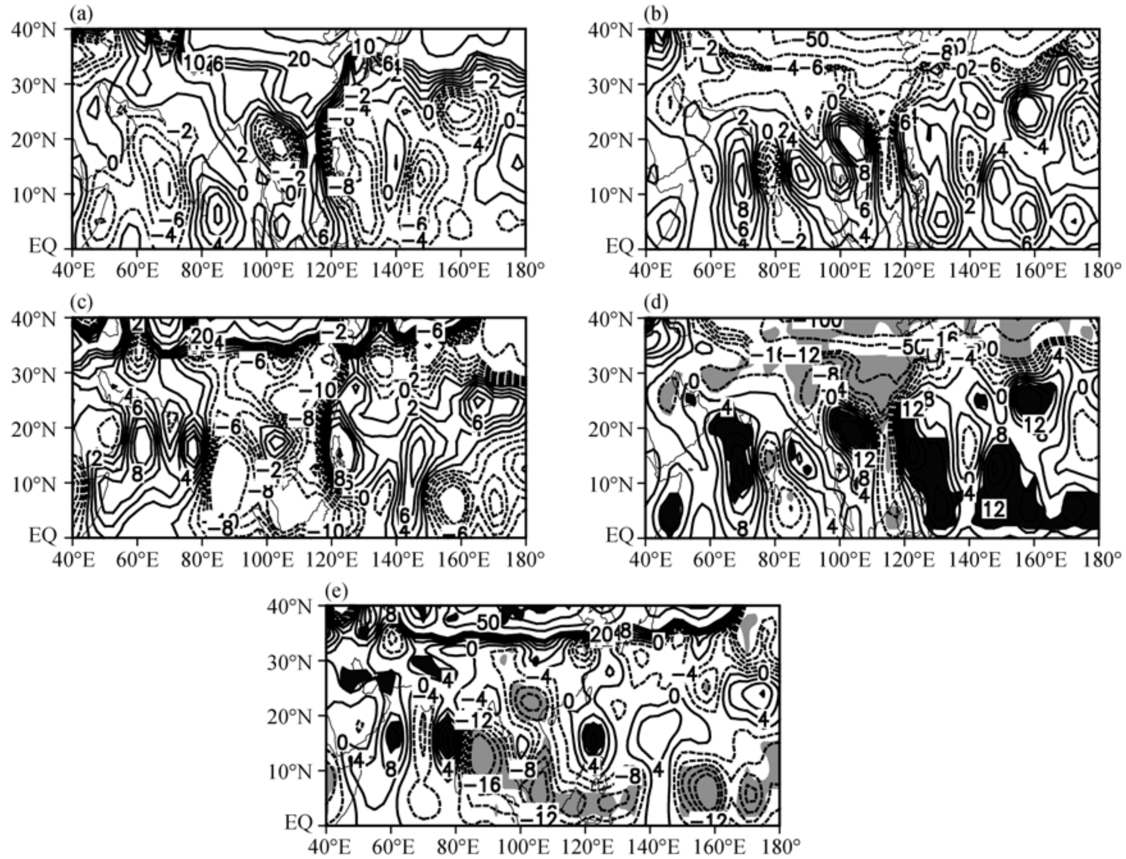

Figure 4 The composite distributions of the STE anomalies of water vapor over Asia in boreal summer (unit: 10^{-9} kg · m^{-2} · s^{-1}) during 1958—1977 (a), 1978—1992 (b), 1993—2001 (c), and composite differences in 1978—1992 minus 1958—1977 (d), 1993—2001 minus 1978—1992 (e). Positive (negative) values significant at 95% confidence level are heavy (light) shaded.

negative ones over the Indian Peninsula to turn to the positive ones. In addition, the cross-tropopause flux anomalies over the WNP have also a distinct change. The other regions except the tropical WNP monsoon region (120°—180°E, 10°—20°N) are dominated by the negative anomalies.

According to the cross-tropopause flux anomalies of water vapor in the three periods, it could be found that the troposphere-to-stratosphere transport of water vapor over BOB-East Asia as well as South China Sea is gradually weakening, while the transport over the Arabian Sea and WNP monsoon region is enhancing. It should be noted that the above results suggest that as important channels of the STE of water vapor, the effect of BOB-East Asia as well as South China Sea is weakening, while the role of the WNP is becoming more and more important. To better understand the interdecadal variations of the water vapor exchange, here we care about two questions. First, what is the reason causing the water vapor exchange to maintain and change on interdecadal scale? Second, what differences have the factors in the various regions? Both questions are worthy of further studying since the interdecadal studies of the water vapor exchange not only have direct importance for better understanding the distribution and variation in the UTLS water vapor, but also may suggest the transport of other components in the atmosphere. The following is aimed at addressing these two questions by studying the characteristics of the atmospheric heat sources and the land-sea thermal contrast on the interdecadal scale from the view of external force of heat source and discussing the relative contribution of land and sea in different periods.

4 Influence of heat source contrast between land and sea on the STE of water vapor over the boreal Asian region

Figures 5(a)—(c) show the mean atmospheric heat source anomalies integrated from surface to 100 hPa in boreal summer in the three periods. Before 1977 (Figure 5(a)), the warm anomalies are located over the regions to north of 20°N and to east of 80°E with high value centers in Somali Peninsula, Indian Peninsula, Southwest China and northern Indo-China Peninsula. The cold anomalies mainly cover West Asia, the Arabian Sea, southern South China Sea, and the WNP to east of the Philippines with a low value center in the Philippine Islands. Although heat source contrast between land and sea is not completely equivalent to the land-sea thermal contrast, both have a close and positive relationship[38]1). Hence, the above distribution helps the BOB and East Asian summer monsoon to stay in the active phase. In the period of 1978—1992 (Figure 5(b)), like the distribution of the water vapor transport, the heat source anomalies over most of the AIPO turn to opposite sign except for the positive anomalies over BOB and Indo-China Peninsula. For instance, the anomalies over the southern South China Sea-the Philippine Islands have changed from the strongest negative center to the strongest positive center, but things are just contrary over the Tibetan Plateau-eastern Chinese mainland. It is evident that the above structure of heat sources weakens the land-sea heat source contrast over East Asia, and leads to weak summer monsoon activity over East Asia and strong summer monsoon activity over the tropical WNP. After 1993 (Figure 5(c)), most regions with positive anomalies appear over the Iran Plateau, and the positive anomalies over the northern Arabian Sea move southeastward. As a result, heat source contrast between land and sea and corresponding ascending flows are further intensified. In addition, Figure 5(c) shows that most regions with negative anomalies are located over the Indo-China Peninsula-central China, while the positive anomalies over the WNP move southward associated with the intensified intertropical convergence zone. From the differences of the summer heat sources shown in Figure 5(d) and (e), it is found that the interdecadal variability over most regions in these three periods is significant. In summary, the interdecadal anomalies of heat sources will cause the sub-systems of the Asian summer monsoon to intensify or weaken by changing heat source contrast between land and sea on the interdecadal scale, and finally results in interdecadal transition of the STE of water vapor over Asia.

Further, we calculate the atmospheric heat source contrast between land and sea during 1958—2001 shown in Figure 6. The heat source contrast between land and sea was gradually increasing from the early 1960s to the early 1970s, and then decreasing to the

1) Zhao P, Zhou X J, Chen L X, et al. Characteristics of monsoon and rainfall over Eastern China and subtropical western North Pacific and associated reasons. Acta Meteorologica Sinica (in Chinese) (no published)

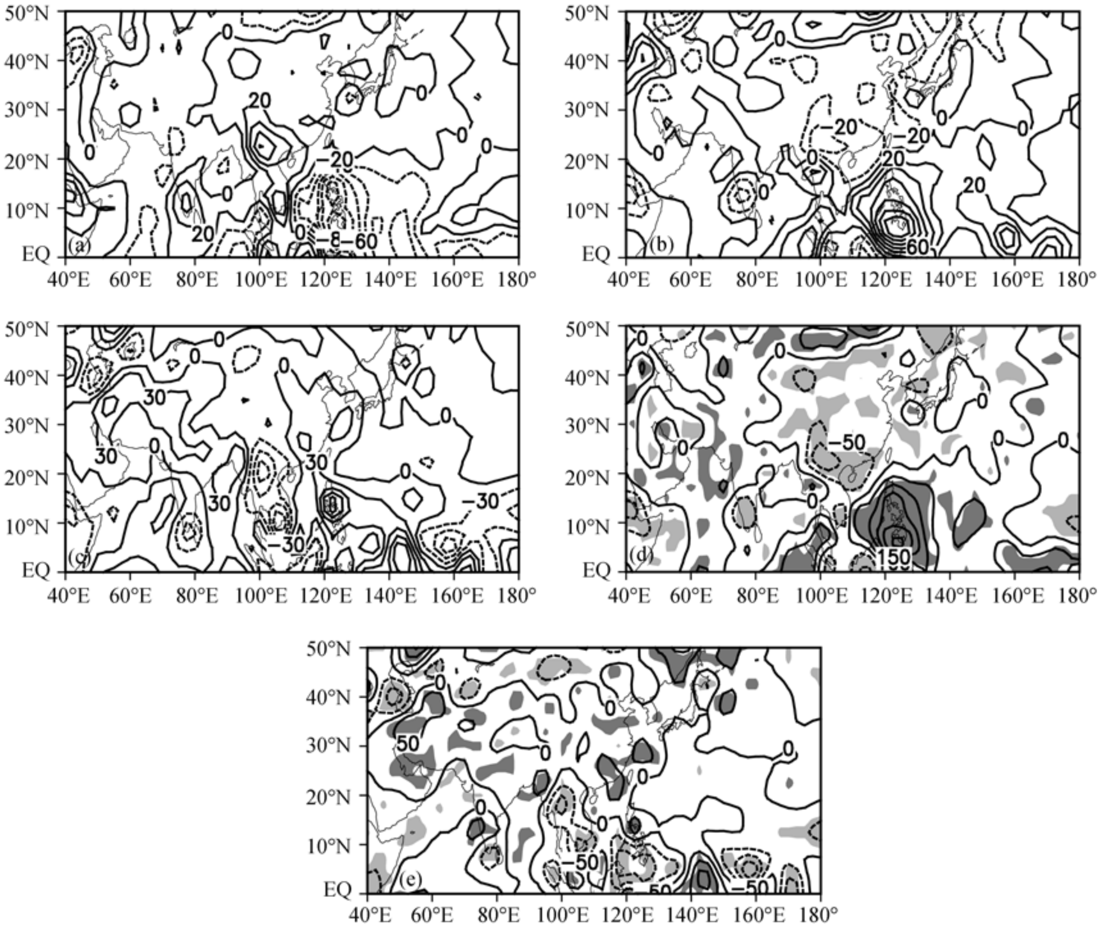

Figure 5 The caption being the same as Figure 4, but for the mean atmospheric heat source anomalies integrated from surface to 100 hPa (Q1, unit: W/m^2). (a) 1958—1977; (b) 1978—1992; (c) 1993—2001; (d) 1978—1992 minus 1958—1977; (e) 1993—2001 minus 1978—1992.

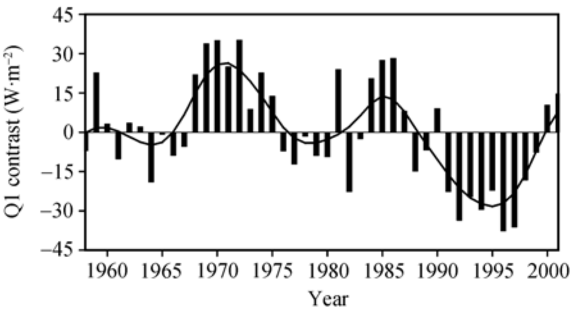

Figure 6 The time series of summer Q1 contrast between land (472 grids) and sea (653 grids) at 30°—140°E and 20°S—40°N (histogram), and its interdecadal component (solid line).

minimum in the late 1970s. After this, the contrast increased again to the maximum in the mid-1980s, but its amplitude was much smaller than that in the early 1970s, and then rapidly reduced to the minimum in the mid-1990s. It is obvious that the land-sea heat source contrast experiences a three-stage interdecadal variability with strong-slight weak-weak trend, and the abrupt points significant at 95% confidence level were located in 1976 and 1990, earlier than the time of interdecadal abrupt of water vapor transport (1977 and 1992). It is therefore concluded that the land-sea synthesis influences the interdecadal variations of the water vapor exchange by changing the Asian summer monsoon. Naturally, the relative importance of land and sea in different periods remains to be clarified.

5 Influence of the atmospheric heat sources over the Tibetan Plateau and the WNP on the STE of water vapor over the boreal Asian region

As indicated in Introduction, the Tibetan Plateau (75°—105°E，25°—40°N) in boreal summer is a remarkable heat source, and its thermal variations play an important role in affecting the Asian monsoon circulation, the formation of ENSO, and the transports of water vapor and mass. Hence, in this study the Tibetan Plateau is regarded as a representative to test the interdecadal variability of heat source over land and its relative importance. On the other hand, the Rossby wave triggered by the thermal convection over the tropical WNP (120°—160°E, 10°—20°N) contributes to the atmospheric general circulation over Asia, and over the WNP, the remarkable interdecadal variability in heat source and water vapor exchange is found. Therefore, the WNP becomes a representative to discuss the interdecadal variability of heat source over sea and its relative importance in this section.

5.1 Interdecadal characteristics of the atmospheric heat sources over the Tibetan Plateau and the tropical WNP

Time series of apparent heat source anomalies over the Tibetan Plateau and the tropical WNP in boreal summer are shown in Figure 7. The most prominent feature is remarkable interdecadal variantions of heat source anomalies over these two regions. As far as heat source over the Tibetan Plateau is concerned (Figure 7(a)), the positive anomalies exist before the mid-1970s and after the early 1990s, and are weaker in recent decade, while the negative anomalies maintain from the mid-1970s to the early 1990s. The heat source over the tropical WNP (Figure 7(b)) exhibits a gradually enhanced trend, and especially intensifies rapidly in recent decade, suggesting that the effects of heat source over the Tibetan Plateau and the cold source over the WNP are strong in the first period, but in the last two periods the heat source over the Tibetan Plateau quickly weakens while that over the WNP rapidly intensifies. The above three-stage interdecadal variations of heat sources are very similar to those of the thermal contrast between land and sea and those of the STE of water vapor, and abrupt change among them is consistent with each other. Is there any relationship between the STE of water vapor and the heat sources on interdecadal scale? An emphasis is put on this question in the following.

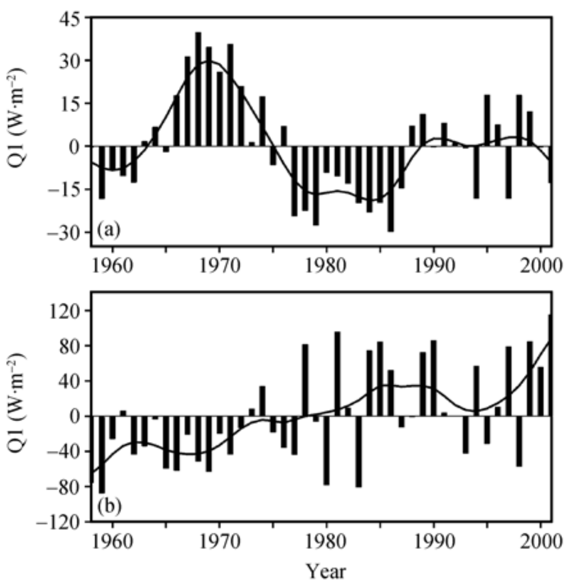

Figure 7 The time series of Q1 (histogram) and its interdecadal component (solid line) over the Tibetan Plateau (75°—105°E, 25°—40°N) (a) and the tropical WNP (120°—160°E, 10°—20°N) (b) in boreal summer.

5.2 Relationship with the STE of water vapor

In order to quantify the interdecadal relationship between heat sources and the STE of water vapor over the Tibetan Plateau and the tropical WNP, we calculate the correlations of the interdecadal components between heat sources over these two regions and the STE of water vapor over Asia in boreal summer, respectively (Figure 8). It should be noted that since the interdecadal components are obtained by 9-point Guassian-type filter, the method of effective number of degrees of freedom by Bretherton et al.[39] is adopted when estimating the significance of correlations and regressions.

It is apparent that both heat sources over the Tibetan Plateau and the tropical WNP are closely related to the water vapor exchange. As far as the heat source over the Tibetan Plateau is concerned (Figure 8(a)), the significant positive correlations are zonally located to the north of 25°N, while the markedly negative correlations distribute intervally over the East Arabian Sea, the East BOB-Indo-China Peninsula, the Philippines and its adjacent sea areas, and the equatorial Pacific to south of 25°N. As far as the heat source over the tropical WNP is concerned (Figure 8(b)), the most prominent feature is that the correlations exhibit a "+ – +" pattern like a wave train from west to east with two boundaries at about 80°E and 120°E. Comparison between Figure 4 and Figure 7 suggests the conclusions as follows: 1) Before 1977, heat source over the Tibetan Plateau and the cold source over the WNP are significantly strong, and their effects on water vapor exchange are consistent. As a result, their synthesis influences the interdecadal variations of water vapor exchange over Asia in this period except BOB-Indo-China Peninsula where the effect of cold source over the WNP is most important. 2) During 1978—1992, negative anomalies over the Tibetan Plateau and positive anomalies over the WNP have also consistent effects on water vapor exchange in Asia except BOB-Indo-China Peninsula where the effect of cold source over the Tibetan Plateau plays a leading role. 3) After 1992, heat source anomalies over the Tibetan Plateau are positive, while those over the WNP persistently intensify with reverse effects in most regions. Further analysis shows that the heat source over the Tibetan Plateau contributes to the STE to north 35°N and equatorial West Pacific, while in other regions, effects of heat source over the WNP are most important.

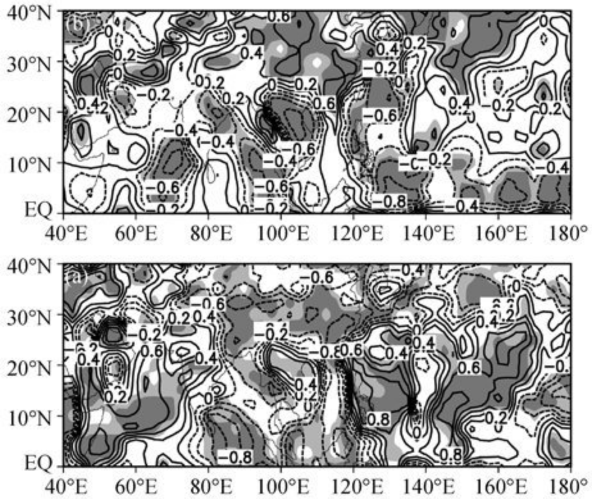

Figure 8 The correlations of the interdecadal components between Q1 over the Tibetan Plateau (a) and the tropical WNP and the STE (b) of water vapor over Asia in boreal summer. Areas in which correlation coefficients are statistically significant at 95% (90%) confidence level are heavy (light) shaded.

The above results suggest that the thermal influences over the Tibetan Plateau and the tropical WNP have been greatly adjusted over the pass 44 years. Their synthesis influences the inter-decadal variations of the water vapor exchange, but their relative importance varies with time and regions. Both contributions are consistent before 1992, but reverse after 1992 when the heat source over the WNP plays a primary role in influencing the across-tropopause water vapor.

5.3 Relationship with the atmospheric circulation

In the previous sections, we have shown that the heat sources over the Tibetan Plateau and the tropical WNP have important influence on the STE of water vapor on the interdecadal scale. Then what is the physical mechanism of such influence? In the following discussion, we will investigate the interdecadal variations of the atmospheric circulation associated with the heat source anomalies.

Figure 9 shows regression patterns of summer 850-hPa wind field against time series of heat sources over the Tibetan Plateau and the tropical WNP on the interdecadal scale, respectively. Figure 9(a) shows that during the period of strong heat source over the Tibetan Plateau, the salient features are the uniform low-level southerly anomalies from South China Sea to East China, and a low-level anticyclone over the WNP, indicating that the subtropical high over the WNP is strongly located much more eastward than its normal position. This circulation pattern matches the active period of typical East Asian summer monsoon, and facilitates convection and precipitation in East China. In addition, we also see the uniform low-level southwesterly anomalies over the whole BOB, the uniform low-level northerly anomalies over the Arabian Sea, and weaker Somali Cross-Equatorial Flow. It suggests that South Asian summer monsoon is active in the east and inactive in the west, which favors the STE of water vapor in the east and restrains that in the west. During the period of strong heat source over the WNP (Figure 9(b)), the atmospheric circulation patterns are nearly contrary to those in Figure 9(a). The uniform low-level northerly anomalies are located over East China-South China Sea, North BOB-southern Indo-China Peninsula, implying that East Asian summer monsoon and BOB summer monsoon are inactive in this period when water vapor is reduced and convection is weak. The uniform low-level southwesterly anomalies are located over the Arabian Sea, which indicates that monsoon over this region is active. In addition, an anomalous anticyclone prevails over the WNP. It means that in this period the subtropical high over the WNP is weaker than normal, and hence favors upward motion. It should be noted that the confluence of easterlies from anomalous cyclone and anomalous westerlies from the equator intensifies the activity of convergence zone over this region.

We now turn our attention to the regression patterns for 500-hPa anomalous geopotential height field shown in Figure 10. At first glance, the regression patterns against time series of heat sources over the Tibentan Plateau and the tropical WNP on the interdecadal scale are out of phase, with the most significant regions over the tropics and southern East Asia to south 20°N. When heat source over the Tibetan Plateau is strong, it is characterized by the strong negative anomalies over most Asian regions, which intensifies the tough in middle and high latitudes and the continental low. In East Asia, this collocation promotes the southward movement of cold

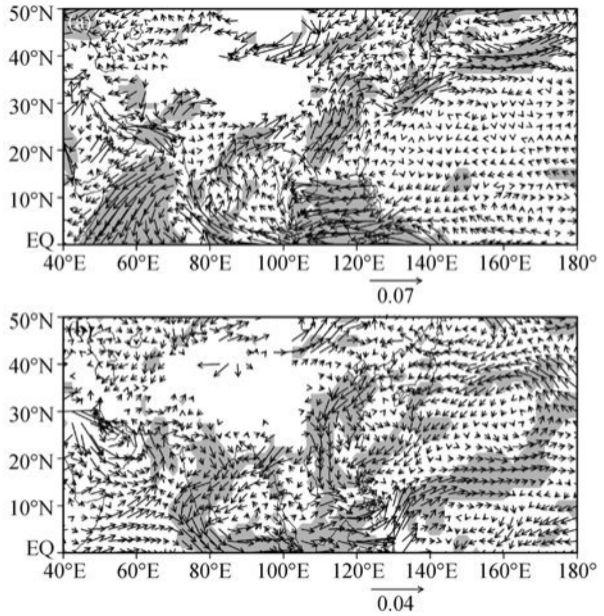

Figure 9 Regression patterns of summer 850-hPa wind field (unit: m/s) against the time series of Q1 over the Tibetan Plateau (a) and the tropical WNP (b) on the interdecadal scale. Areas in which regression coefficients are statistically significant at 95% confidence level are shaded.

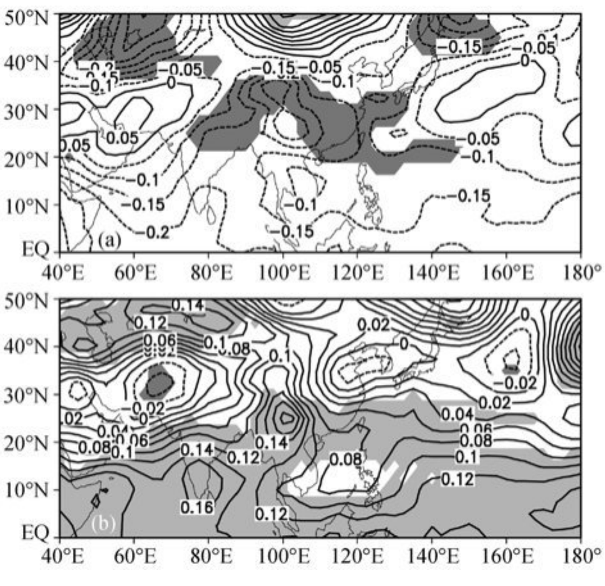

Figure 10 The caption being the same as Figure 9, but for 500-hPa anomalous geopotential height field (unit: 10 gpm). Negative (positive) values significant at 95% confidence level are heavy (light) shaded.

air from north and the northward advance of warm and moist air from south so that convection over the front and water vapor transport intensify. Similarly, in BOB, this collocation is also helpful to the transport of warm and moist air flows from southwest. However, the pattern when heat source over the WNP is strong is contrary to the above one. Figure 10(b) shows that: 1) the subtropical high over the WNP extends westwards which damps the northward movement of warm and moist air from south; 2) the Indian low weakens which restrains the northward advance of warm and moist air from southwest over BOB.

Since the STE of water vapor is closely related to the distribution and variation of tropospheric water vapor, the regression patterns for water vapor flux divergence integrated from surface to 300 hPa are shown in Figure 11. When heat source over the Tibetan Plateau is strong, convergence areas of water vapor flux are located in BOB, East Asian continent and to the west of Japan, corresponding to active summer monsoon over these regions, while divergence areas mainly cover the Arabian Sea, South China Sea-ocean to east of the Philippines, associated with inactive summer monsoon over these regions. In comparison with Figure 11(a), when heat source over the WNP is strong, distribution of water vapor flux divergence is different. It is found that convergence areas of water vapor flux control most Asian regions, while divergence areas are just situated over eastern Asian continent and Indian Peninsula-west BOB. This indicates that there are more water vapor and stronger upward transport over most Asian regions in this period, thus the transport from the troposphere to the stratosphere intensifies.

Next, let us examine high-level vertical motion field. Figure 12 shows the regression patterns for 100-hPa vertical motion field. Similarly, 100-hPa vertical motion

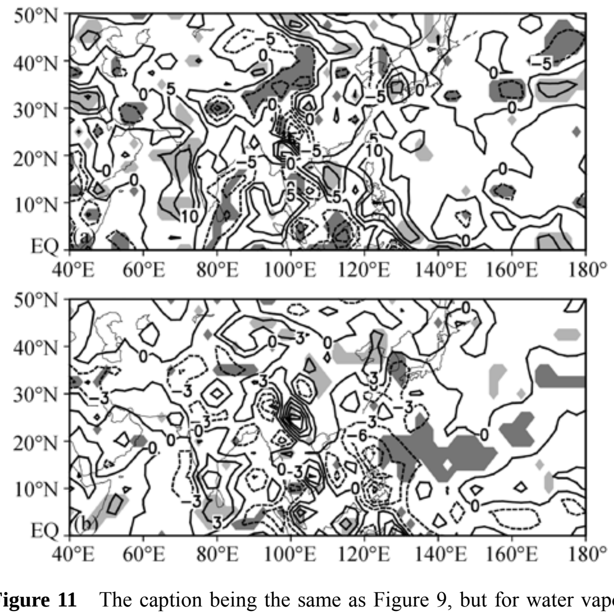

Figure 11 The caption being the same as Figure 9, but for water vapor flux divergence integrated from surface to 300 hPa (unit: 10^{-10} $kg \cdot m^{-2} \cdot s^{-1}$). Convergence (divergence) areas significant at 95% confidence level are heavy (light) shaded.

field associated with heat source over the Tibetan Plateau tends to be out of phase with that associated with the WNP on interdecadal scale. During the period of strong heat source over the Tibetan Plateau (Figure 12(a)), anomalous upward motion is evident over the Arabian Peninsula, Indian Peninsula, China mainland-South China Sea-south Indo-China Peninsula, and northern West Pacific, while anomalous downward motion is over the eastern Arabian Sea, North Indo-China Peninsula, the Philippines and ocean to east of it. When heat source over the WNP is strong (Figure 12(b)), the anomalous vertical wind field is nearly opposite to that in Figure 12(a). It is clear that the anomalous sinking areas are basically located to north of 25°N, while the anomalous upward motion occurs over the Arabian Sea, the Philippines and ocean to east of it. It is apparent that the regression patterns for 100-hPa vertical motion field are similar to the relation patterns shown in Figure 8, indicating that the variations of heat source trigger the anomalous vertical motion and monsoon circulation, and thus intensify the cross-tropopause transport of water vapor.

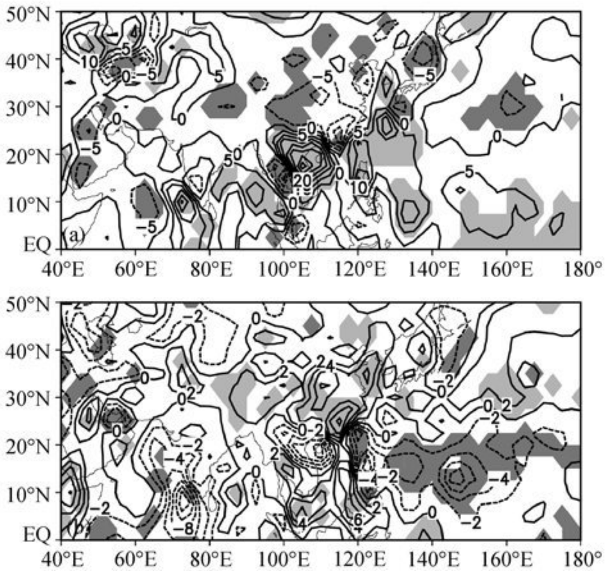

Figure 12 The caption being the same as Figure 9, but for 100-hPa vertical motion field (unit: 10^{-5} Pa/s). Negative (positive) areas significant at 95% confidence level are heavy (light) shaded.

In summary, the heat sources over the Tibetan Plateau and the WNP have important influence on the STE of water vapor over Asia on interdecadal scale, but their contributions vary with time and regions. Both contributions are consistent before 1992, but reverse after 1992 when the heat source over the WNP plays a main role in influencing the across-tropopause water vapor. The mechanism influencing the STE of water vapor is concluded as follows: The heat source anomalies over land and sea in boreal summer cause the anomalies of Asian summer monsoon activities. Impacted by different sub-systems of Asia Asian summer monsoon, the atmospheric circulation and vertical motion in Asia generate the abnormality, and then influence the variations of water vapor exchange.

6 Conclusion and discussion

The STE of water vapor in the Northern Hemisphere has been diagnosed with the Wei method. Particularly, focusing on Asia in boreal summer, inter-decadal variations have been studied. 44-year climatology of the STE of water vapor in the Northern Hemisphere has shown that the STE of water vapor is characterized by transport from troposphere to stratosphere in the tropics and from stratosphere to troposphere in the extra-tropics. It should be noted that the joining area of the AIPO becomes the main channel of the upward transport of water vapor across the tropopause due to the Asian summer monsoon onset. This channel can persistently convey the abundant water vapor into the stratosphere and then influence the distribution and variation of the stratospheric water vapor. In addition, it is found from the distribution of ground emission sources presented in ref. [40] that BOB-northeast Indian Peninsula is a highly polluted area across which the emission is likely to enter the stratosphere. Therefore, a focus is put on the Asian region in this study.

EOF analysis and abrupt change test method show that there are significant inter-decadal variations in the water vapor exchange over Asia, and the recent 44 years can be divided into the following three periods: 1958—1977, 1978—1992 and 1993—2001. In these three periods, as important channels of the water vapor exchange, the effects of BOB-East Asian continent as well as South China Sea are gradually weakening, while the role of the WNP becomes more and more important.

It is clear from our results that the atmospheric heat sources over land and sea are characterized by the prominent interdecadal variability, consistent with the interdecadal transition of STE of water vapor. Further, the Tibetan Plateau and the tropical WNP are regarded as representatives of land and sea, respectively, to investigate the relative importance of continental and ocean heat sources in the STE of water vapor. We conclude that the thermal influences over the Tibetan Plateau and the WNP have been greatly adjusted over the pass 44 years. Their synthesis influences the inter-decadal variations of the water vapor exchange by changing the Asian summer monsoon, but their roles vary with time and regions. Both contributions are consistent before 1992, but reverse after 1992 when the heat source over the WNP plays a main role in influencing the across-tropopause water vapor. The inter-decadal variations of heat sources over the Tibetan Plateau and the tropical WNP are therefore thought to be the two main factors in determining the inter-decadal variations of water vapor exchange.

At the end of last century, Oltmans and Hofmann[41] first noticed the increasing trend in the stratospheric water vapor by using the radiosonde data. Do the interdecadal variations of STE of water vapor have influence on the long-time trend of the stratospheric water vapor? If yes, which region on earth plays the main role? The above questions are so important that they will become a key in future research.

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1 Brewer A W. Evidence for a world circulation provided by the measurements of helium and water vapour distribution in the stratosphere. Q J R Meteorol Soc, 1949, 75: 351—363

2 Holton J R, Haynes P H, McIntyre M E, et al. Stratosphere- troposphere exchange. Rev Geophys, 1995, 33: 403—439

3 Newell R E, Gould-Stewart S. A stratospheric fountain? J Atmos Sci, 1981, 38: 2789—2796

4 Rosenlof K. How water enters the stratosphere. Science, 2003, 302: 1691—1692

5 Chen P. Isentropic cross-tropopause mass exchange in the extratropics. J Geophys Res, 1995, 100: 16661—16673

6 Dethof A, O'Neill A, Slingo J M, et al. A mechanism for moistening the lower stratosphere involving the Asian summer monsoon. Q J R Meteorol Soc, 1999, 125: 1079—1106

7 Cong C H, Li W L, Zhou X J. Mass exchange between stratosphere and troposphere over the Tibetan Plateau and its surroundings. Chin Sci Bull, 2002, 47 (6): 508—512

8 Yang J. Study on the process of stratosphere-troposphere exchange over eastern Asia. Doctor Dissertation For Doctorial Degree (in Chinese). Beijing: Institute of Atmospheric Physics of Chinese Academy of Sciences, 2002

9 Bannister R N, O'Neill A, Gregory A R, et al. The role of the S. E. Asian Monsoon and other seasonal features in creating the 'tape recorder' signal in the unified model. Q J R Meteorol Soc, 2003, 130: 1531—1554, doi:10.1256/qj.03.106

10 Gettelman A, Kinnison D E, Dunkerton T J, et al. The impact of monsoon circulations on the upper troposphere-lower stratosphere. J Geophys Res, 2004, 109: D22101, doi:10.1029/2004JD004878

11 Fu R, Hu Y, Wright J S, et al. Short circuit of water vapor and polluted air to the global stratosphere by convective transport over the Tibetan Plateau. Proc Nat Acad Sci, 2006, 103: 5664—5669

12 Zhan R, Li J, Gettelman A. Intraseasonal variations of upper tropospheric water vapor in Asian monsoon region. Atmos Chem Phys Discuss, 2006, 6: 8069—8095

13 Guo D, Lu D R, Sun Z B. Seasonal variation of global stratosphere-troposphere mass exchange. Prog Nat Sci, 2007, 17(12): 1466—1475

14 Tao S Y, Chen L X. The structure of general circulation over continent of Asia in summer. Acta Meteorologica Sinica (in Chinese), 1957, 28(2): 234—246

15 Zeng Q C, Li J P. Interactions between the Northern and Southern Hemispheric Atmospheres and the Essence of Monsoon. Chin J Atmos Sci (in Chinese), 2002, 26: 433—448

16 Ye D Z, Wu G X. The role of the heat source of the Tibetan Plateau in the general circulation. Met Atmos Phys, 1998, 67: 181—198

17 Zhao P, Chen L X. Climatic features of atmospheric heat source/sink over the Qinghai-Xizang Plateau in 35 years and its relation to rainfall in China. Sci China Ser D-Earth Sci, 2001, 44(9): 858—864

18 Xu X D, Zhou M Y, Chen J Y, et al. A comprehensive physical pattern of land-air dynamic and thermal structure on the Qinghai-Xizang Plateau. Sci China Ser D-Earth Sci, 2002, 49(2): 181—188

19 Chen L X, Shmidt F, Li W. Characteristics of the atmospheric heat source and moisture sink over the Qinghai–Tibetan Plateau during the second Tipex of summer 1998 and their impact on surrounding monsoon. Meteor Atmos Phy, 2003, 83: 1—18

20 Chen H B, Bian J C, Lu D R. Advances and prospects in the study of stratosphere-troposphere exchange. Chin J Atmos Sci (in Chinese), 2006, 30: 813–820

21 Nitta T. Convective activities in the tropical western Pacific their impact on the Northern Hemisphere summer circulation. J Meteor Soc Japan, 1987, 64: 373—390

22 Huang R H, Sun F Y. Impact of the tropical western Pacific on the East Asian summer monsoon. J Meteor Soc Japan, 1992, 70: 243—256

23 Lu R Y, Ryu C S, Dong B W. Associations between the Western North Pacific Monsoon and the South China Sea monsoon. Adv Atmos Sci, 2002, 19: 12—24

24 Chou C, Tu J Y, Yu J Y. Interannual Variability of the Western North Pacific Summer Monsoon: Differences between ENSO and Non-ENSO Years. J Clim, 2003, 16: 2275—2287

25 Wei M Y. A new formulation of the exchange of mass and trace constituents between the stratosphere and troposphere. J Atmos Sci, 1987, 44 (20): 3079—3086

26 Siegmund P C, van Velthoven P F J, Kelder H. Cross-tropopause transport in the extratropical northern winter hemisphere, diagnosed from ECMWF data. Q J R Meteorol Soc, 1996, 122: 1921—1941

27 Lamarque J F, Hess P G. Cross-tropopause mass exchange and po-

tential vorticity budget in a simulated tropopause folding. J Atmos Sci, 1994, 51: 2246—2269
28 Spaete P, Johnson D R, Schaack T K. Stratospheric-tropospheric mass exchange during the President's Day strom. Mon Wea Rev, 1994, 122: 424—439
29 Wirth V, Egger J. Diagnosing extratropical synoptic-scale stratosphere-troposphere exchange: A case study. Q J R Meteorol Soc, 1999, 125: 635—655
30 Gettelman A, Sobel A H. Direct diagnoses of stratosphere-troposphere Exchange. J Atmos Sci, 2000, 57(1): 3—16
31 Wang W G, Fan W X, Jing J D, et al. The characteristics of spatial-temporal variations of the global cross-tropopause mass flux. Plateau Meteorol (in Chinese), 2007, 26(5): 910—920
32 WMO. Definition of the tropopause. WMO Bull 6, 1957, 1—136
33 Yanai M, Esbensen S, Chu J H. Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets. J Atmos Sci, 1973, 30: 611—627
34 Ding Y H. Research Methods of Synoptic Dynamics. Beijing: Science Press, 1989. 146
35 MAXOBEP 3 M. Climatology of the Tropopause (in Russian). Translated by Zhang G Y, Liao S F. Beijing: China Meteorological Press, 1988. 1—303
36 Wu G X, Li J P, Zhou T J, et al. The key region affecting the short-term climate variations in China: The joining area of Asia and Indian-Pacific Ocean. Adv Earth Sci (in Chinese), 2006, 21(11): 1109—1118
37 North G, Bell T, Cahalan R, et al. Sampling errors in the estimation of empirical orthogonal function. Mon Weather Rev, 1982, 110: 699—706
38 Li C, Yanai M. The onset and interannual variability of the Asian summer monsoon in relation to land-sea thermal contrast. J Clim, 1996, 9 (2): 358—375
39 Bretherton C S, Widmann M, Dymnikov V P, et al. Effective number of degrees of freedom of a spatial field. J Clim, 1999, 12: 1990—2009
40 Kato N, Akimoto H. Anthropogenic emissions of SO_2 and NO_X in Asia: Emission inventories. Atmos Environ, 1992, 26A: 2997—3017
41 Oltmans S J, Hofmann D J. Increase in lower-stratospheric water vapour at a mid-latitude northern hemisphere site from 1981—1994. Nature, 1995, 374: 146—149