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A CASE STUDY OF THE UNCORRELATED RELATIONSHIP BETWEEN TROPICAL TROPOPAUSE TEMPERATURE ANOMALIES AND STRATOSPHERIC WATER VAPOR ANOMALIES

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Abstract: Using the measurements from the Halogen Occultation Experiment (HALOE) and the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim reanalysis data for the period 1994–2005, we analyzed the relationship between tropical tropopause temperature anomalies and stratospheric water vapor anomalies. It is found that tropical tropopause temperature is correlated with stratospheric water vapor, i.e., an anomalously high (low) tropical tropopause temperature corresponds to anomalously high (low) stratospheric water vapor during the period 1994–2005, except for 1996. The occurrence frequency and strength of deep convective activity during the 'mismatched' months is less and weaker than that during the 'matched' months in 1996. However, the instantaneous intensity of four short periods of deep convective activity, caused by strong surface cyclones and high sea surface temperatures, are greater during the 'mismatched' months than during the 'matched' months. Water vapor is transported from the lower troposphere to the lower stratosphere through a strong tropical upwelling, leading to an increase in stratospheric water vapor. On the other hand, deep convective activity can lift the tropopause and cool its temperature. In short, the key factor responsible for the poor correlation between tropical tropopause temperature and stratospheric water vapor in 1996 is the instantaneous strong deep convective activity. In addition, an anomalously strong Brewer-Dobson circulation brings more water vapor into the stratosphere during the 'mismatched' months in 1996, and this exacerbates the poor correlation between tropical tropopause temperature and stratospheric water vapor.

Key words: tropical tropopause temperature; stratospheric water vapor; deep convective activity; Brewer-Dobson circulation; ENSO

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1 INTRODUCTION

As one of the most important trace gases in the stratosphere, water vapor plays a crucial role in the planetary radiation budget, energy balance and stratospheric chemistry (Held and Brian [1]; Forster and Shine [2]; Zhang et al.[3]). Solomon et al. pointed out that the concentration of stratospheric water vapor decreased by about 10% since 2000, which has led to a slow-down of the increase rate of surface warming over the period 2000–2009 by about 25% when compared with that caused only by carbon dioxide and other greenhouse gases[4]. Several studies have shown that the concentrations and spatial distribution of stratospheric water vapor are dominated mainly by three factors: effective freeze-drying of the air associated with the cold tropical tropopause temperature (Holton et al. [5]; Andrews et al. [6]), the transport effect of stratospheric water vapor through the large-scale Brewer–Dobson circulation (BDC) (Brewer[7]) and methane oxidation in the upper stratosphere (Texier et al. [8]; Tian and Chipperfield[9]). In recent years, the development of new observation methods and detection techniques, such as ground-based and sounding observations, has led to the collection of relatively rich water-vapor datasets from the middle and upper atmosphere. Consequently, the changing distribution of stratospheric water vapor has begun to draw a great deal of attention and many new perspectives have emerged on this topic (Shindell [10]; Stenke and Grewe[11]; Chipperfield[12]). It has been shown

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that stratospheric water vapor exhibited an increasing trend over the past half-century (Kely et al.[13]; Rosenlof et al. [14]). By using sounding observations and the Halogen Occultation Experiment (HALOE) satellite measurements, Oltmans and Holmann, Oltmans et al. and Nedoluha et al. confirmed that water vapor amounts in the lower stratosphere have increased at a rate of 1% /yr since 1980, and that the increase rate is fastest during the period 1993–1997 [15–18]. On the one hand, the Intergovernmental Panel on Climate Change (IPCC) reported that although methane oxidation is one of the most important sources of stratospheric water vapor, the positive long-term trend in stratospheric water vapor over the past half-century cannot be explained solely by the increase in methane, which can only account for an average annual growth rate of water vapor of about 24%–32%(Intergovernmental Panel on Climate Change, 2013 [19]). On the other hand, Stratosphere-Troposphere Processes And their Role in Climate (SPARC) showed that changes in the amount of stratospheric water vapor that is transported from the troposphere make the major contribution to the positive long-term trend in stratospheric water vapor over this period (Kley et al.[13]). As early as 1946, Dobson et al. found that the saturated mixing ratio of water vapor in the tropical tropopause is consistent with that in the stratosphere [20]. Subsequently, there has been growing evidence showing that tropical tropopause temperature is the key factor controlling stratospheric water vapor (Mote et al.[21]; Fueglistaler et al. [22]). When air enters the stratosphere from the troposphere in the tropics, a cold tropical tropopause would induce very low water-vapor mixing ratios throughout the stratosphere by effective freeze-drying of the air (Seidel et al.[23]; Rosenlof et al.[24]). In terms of the interannual timescale of stratospheric water vapor, Tian and Chipperfield used a coupled chemistry-climate model (CCM) to demonstrate that the trend of +0.44 K/decade in the temperature at 100 hPa has contributed up to 70% (+35 ppbv/yr) of the water vapor trend in the lower stratosphere [25]. Rosenlof et al. indicated that the positive trend in stratospheric water vapor over the past half-century is in close agreement with the warming trend of the tropical tropopause temperature [14]. Consequently, atmospheric processes which have an influence on the tropical tropopause temperature (i.e., deep convective activity in the tropics and the tropical quasi-biennial oscillation, QBO), can affect the amount of water vapor entering the stratosphere (Lu et al. [26]; Tian et al.[27]).

However, it should be pointed out that despite the close relationship between tropical tropopause temperature and stratospheric water vapor, there is also a scenario in which an anomalously high (low) tropical tropopause temperature coincides with anomalously low (high) stratospheric water vapor in certain years, and we refer to this phenomenon as an uncorrelated relationship (Gettelman et al.[28]). Fueglistaler found this phenomenon in 1991, specifically, there is a decrease in tropical tropopause temperature but an increase in stratospheric water vapor [29]. They also pointed out that the possible cause of this uncorrelated relationship is that stronger tropical upwelling caused by the eruption of Mt. Pinatubo in June 1991 led to the colder tropaupose temperature, and the eruption injected water vapor directly into the stratosphere. Consequently, stratospheric water vapor is affected less by the tropical tropopause temperature, resulting in the development of an uncorrelated relationship. It would be worthwhile to examine whether other years without volcanic eruptions also show this uncorrelated relationship between tropical tropopause temperature and stratospheric water vapor, such as that observed in 1991. If so, what is the cause of this uncorrelated relationship? As mentioned above, apart from the tropical tropopause temperature, the large-scale BDC and tropical deep convective activity can result in water vapor changes via anomalous dynamical transport. Under such conditions, an anomalously high (low) tropical tropopause temperature may correspond to anomalously low (high) stratospheric water vapor.

By using the measurements from HALOE and the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim reanalysis data for the period 1994–2005, we find a further example of an obvious uncorrelated relationship between tropical tropopause temperature and stratospheric water vapor. Further analysis is required on the cause of this uncorrelated relationship.

2 DATA AND METHODS

The HALOE measurement data were collected by NASA's Upper Atmosphere Research Satellite (UARS) launched in September 1991, with a global coverage from 80° N to 80° S and 271 vertical pressure levels from the surface to 10^{-6} hPa (Russell et al. [30]). The tropical tropopause temperature presented a significant increasing trend after the eruption of Mt. Pinatubo in June 1991, but tended to decrease in the subsequent years (Xin and Tian[31]; Xu[32]). Such volcanic eruptions ejected a large amount of volcanic aerosols and other substances into the stratosphere, which has a significant impact on the stratospheric trace gases (Zheng [33]). To avoid the influence of the eruption of Mt. Pinatubo on the tropical tropopause temperature and stratospheric water vapor over the following three years, we use the monthly water vapor observations derived from HALOE (http://haloe.gats-inc.com/download/index.php) for the period from 1994 to 2005. In addition, we process the source HALOE datasets for convenience and analyze the gridded data for 12 years at a horizontal resolution of 3° latitude × 3° longitude and with 271 pressure levels (Zheng and Shi [34]). Earlier studies showed that ECMWF Interim (ERA-Interim) temperatures are broadly consistent with those of other data sets (Monge et al.[35]; Rienecker et al.[36]; Saha et al.[37]), and their water vapor datasets are in good coincidence with those of Microwave Limb Sounder (MLS) in the spatial distribution (Xie et al.[38]). Hence, daily datasets from ERA-Interim which are available at 00:00, 06:00, 12:00, 18:00 UTC every day (the same below) with a horizontal resolution of 1° latitude × 1° longitude and 37 vertical pressure levels together with the HALOE satellite data are used to analyze the thermodynamic and dynamical cause of the uncorrelated relationship between the tropical tropopause temperature and stratospheric water vapor.

We also use the sea surface temperature (SST) and outgoing long-wave radiation (OLR) datasets from National Centers for Environmental Prediction (NCEP), which have a horizontal resolution of 2.5° latitude × 2.5° longitude and 17 vertical pressure levels. We compare the National Oceanic and Atmospheric Administration (NOAA, Extended Reconstructed SST-ERSST.v3) SST dataset with that from NCEP. Note that OLR has been shown to be a good indicator of deep convection in the tropics. To diagnose the transport effect of deep convective activity, high cloud data from ECMWF is applied as a cross-analysis with the OLR data. By comparing the cloud data from ECMWF with the International Satellite Cloud Climatology Project (ISCCP), Jakob found that cloud data from ECMWF is able to capture the major features of the interannual variability of cloud [39]. Previous studies have shown that the Cold Point Tropopause (CPT) has a closer relationship with stratospheric water vapor in the tropics, compared with the thermal tropopause temperature defined by the World Meteorological Organization (WMO) (Randel et al.[40]). Thus, CPT temperature (CPTT) is used to analyze its relationship with stratospheric water vapor in this study. Simmons et al. indicated that specific humidity derived from ECMWF data shows good consistency with other independent water vapor data when analyzing the water vapor near the tropopause [41], thus, we use specific humidity to represent stratospheric water vapor. In the following, unless otherwise stated, all anomalies are defined as deviations of a given variable from its seasonal cycle.

3 CASE STUDY

Figure 1 shows the time series of tropical CPTT and 70 hPa water vapor anomalies (averaged over 10.5° N–10.5° S) for the period 1994–2005 after the removal of linear trends. From the annual mean time series (Fig.1a), we can see that the interannual variations in water vapor derived from the ERA-Interim data are in agreement with those obtained from HALOE. To be specific, the tropical CPTT is closely correlated with 70 hPa water vapor during most of the period 1994–2005. However, note that there is a negative anomaly of CPTT in 1996, which corresponds to the positive anomaly of stratospheric water vapor rather than the negative water vapor anomaly. The situation in 1999 is the opposite to that in 1996, with the anomalously high tropical CPTT corresponding to anomalously low stratospheric water vapor instead of high water vapor. This indicates an uncorrelated relationship between CPTT and stratospheric water vapor in these two years. Fig.1b shows monthly mean variations in CPTT and 70 hPa water vapor anomalies. Positive CPTT anomalies are evident in June, July and August 1999, and these positive CPTT anomalies are strongest over the period 1994–2005. However, there are negative anomalies in stratospheric water vapor, forming the uncorrelated relationship with the positive CPTT anomalies. Further analysis of the Oceanic Nino Index (ONI) shows that the ONI is less than –0.5℃ in June, July and August 1999, indicating the occurrence of a strong La Niña event during this period. It is well known that El Niño-Southern Oscillation (ENSO) has a significant impact on variations in CPTT by affecting the deep convective activity (Zhou et al.[42]; Xie et al.[43]). In general, during La Niña events, SST decreases over the tropical eastern Pacific and deep convective activity is relatively weak there, corresponding to the colder CPT (Zhou et al.[42]). Stratospheric water vapor is characterized mainly by dryness during La Niña events (Scaife et al.[44]). Hence, the presence of the La Niña event is the main factor causing the uncorrelated relationship between CPTT and stratospheric water vapor in 1999. It is worth noting that the La Niña event in 1999 extends into 2000. A comparison of the anomalous SST, OLR and vertical velocity of the BDC ($\bar{\omega}^*$) between 1999 and 2000 (not shown here) shows a negatively anomalous SST, positively anomalous OLR and positively anomalous $\bar{\omega}^*$ almost everywhere in the tropics (10.5° N–10.5° S) in 1999, which lead to weakened convective activity and tropical upwelling. Weakened convective activity and tropical upwelling act to lower the tropical tropopause height and increase the CPTT, but reduce the transport of water vapor into the lower stratosphere. In 2000, SST is low over the Eastern and Central Pacific, but high over the Western Pacific, and the OLR signals are of the opposite signs to SST, indicating that SST and OLR change only weakly throughout the tropics, and the change in $\bar{\omega}^*$ is also subtle. Accordingly, there is an uncorrelated relationship between CPTT and stratospheric water vapor in 1999, but not in 2000. Furthermore, there are the westerly phase of the QBO in June, July and August 1999. Anomalous westerly wind shear during the westerly phase of the QBO, which descends from the upper stratosphere to the lower stratosphere, induces an anomalous sinking motion that leads to a warmer and lower tropopause in the tropics (Reid and Gage[45]; Collimore et al.[46]). In addition, under the westerly phase of the QBO, water vapor shows a negative anomaly in the lower stratosphere (Shi et al.[47]), further intensifying the uncorrelated relationship in 1999. Hence, no further analysis will be presented for the reason of the uncorrelated relationship in 1999, and the goal of this paper is to explore the uncorrelated relationship in 1996.

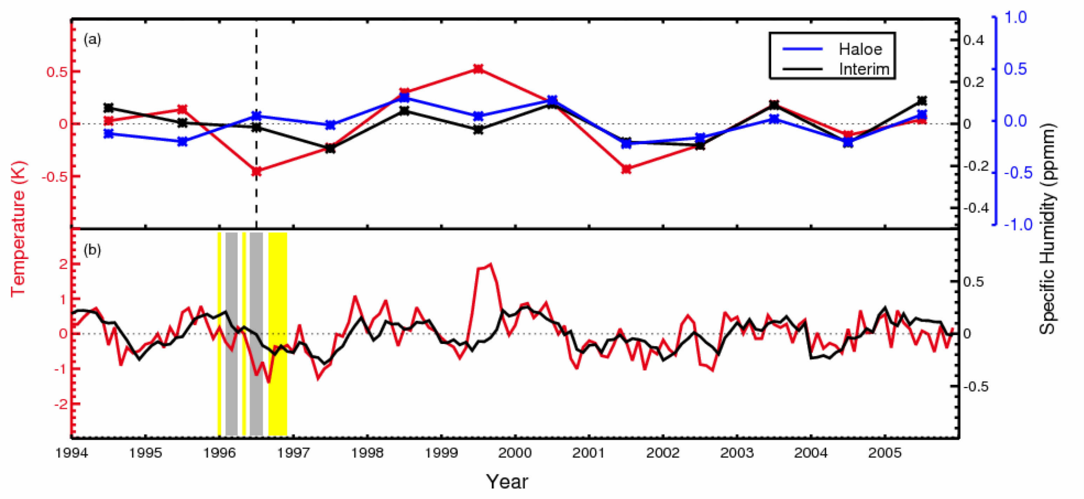

Figure 1. Time series of the (a) annual mean and (b) monthly mean tropical CPTT anomalies and water vapor anomalies at 70 hPa (averaged over 10.5°N–10.5°S) for the period 1994–2005. Straight line represents CPTT. Dot-dash line and dashed line represent water vapor derived from the ERA-Interim reanalysis and HALOE data, respectively. ‘Matched’ months in 1996 are denoted by the yellow area and ‘mismatched’ months by the light grey area.

In Fig.1b, for the ‘mismatched’ months of February, March, April, June, July and August 1996 between CPTT and stratospheric water vapor are indicated by the light grey area. Note that there are significant negative anomalies of CPTT in June, July and August 1996, corresponding to the slightly decreased stratospheric water vapor. The magnitude of the reduction in amplitude varies between CPTT and stratospheric water vapor, which leads to the formation of their uncorrelated relationship. It seems likely that there is an increase of stratospheric water vapor in February, March and April 1996, and this increase is maintained over the next few months, thereby partially offsetting the freeze-drying of the air by the significantly colder CPTT in June, July and August 1996. Therefore, we focused mainly on the uncorrelated relationship between CPTT and stratospheric water vapor in February, March and April 1996. To highlight the characteristics of the CPTT and stratospheric water-vapor distribution during the ‘mismatched’ months, we defined the remaining months in 1996 as the ‘matched’ months (yellow area in Fig.1b), and these are used as a comparison with the ‘mismatched’ months.

Figures 2a and 2b show longitude-height cross-sections of temperature and water vapor anomalies during the ‘mismatched’ months of 1996. There are evident negative anomalies of CPTT and positive stratospheric water vapor anomalies (above 100 hPa), which highlight their uncorrelated relationship. Fig. 2b shows large positive water vapor anomalies developing in the upper troposphere over 0–30°E, 70–100°E, 120–150° E, 90 –75° W and 25° W –0°, among which the positive water anomaly vapor stretches directly into the lower stratosphere over the regions 70–100°E and 120–150°E. This indicates that the positive anomaly in water vapor in the lower stratosphere is closely related to the transport processes of water vapor in these two areas. However, positive anomalies in water vapor over the other areas only stretch from 300 to 150 hPa, implying that an increase in stratospheric water vapor is not caused by the positive water vapor anomaly over these regions. The distribution of temperature and water vapor anomalies during the ‘matched’ months (Figs.2c and d) confirms the close correlation between CPTT and stratospheric water vapor. Note that large positive water vapor anomalies develop in the upper troposphere over 100–150°E and 60–30°W, but stretch only to 150 hPa, and not directly to the lower stratosphere. Therefore, there is a positive correlation between CPTT and stratospheric water vapor during the ‘matched’ months.

Figure 3 shows longitude-height cross-sections of anomalous water vapor and vertical velocity (Figs.3a and c, respectively), and the latitudinal distribution of OLR and the OLR anomaly (Figs.3b and d, respectively). It is apparent that water vapor, vertical velocity, and OLR are in agreement with each other, with the positive (negative) anomalous water vapor regions coinciding with the positive (negative) anomalous vertical velocity and negative (positive) anomalous OLR regions. Specifically, over 70 –100° E and 120 –150° E, the positive water vapor anomaly stretches from the lower troposphere (1,000 hPa) into the lower stratosphere (30 hPa) and corresponds to the positive anomaly in vertical velocity, which represents enhanced vertical rising. Previous studies have referred to OLR<240 W·m^{-2} as being a reliable indicator of deep convective activity (Jury[48]; Kousky[49]). The regions where water vapor is directly transported from the troposphere into the stratosphere are consistent with those having small negative OLR anomalies and large absolute OLR values. During the ‘matched’ months (Figs.3c and d), water vapor is transported from the lower troposphere into the upper troposphere over 100–150 °E and 60–30 °W, corresponding to the positive vertical velocity anomaly and negative OLR anomaly, and OLR is less than 240 W·m^{-2}. A comparison of the water vapor and OLR distributions between 100–150°E and 60–30°W shows that water vapor over 100–150°E is transported higher than that over 60–30°W, and that absolute OLR values over 100–150°E are the greatest seen over the study period. This indicates that strong deep convective activity occurs over 70–100° E and 120–150°E during the ‘mismatched’ months, but over 100–150°E during the ‘matched’ months, and these regions represent the main channels for the transport of water vapor from the lower troposphere into the upper troposphere.

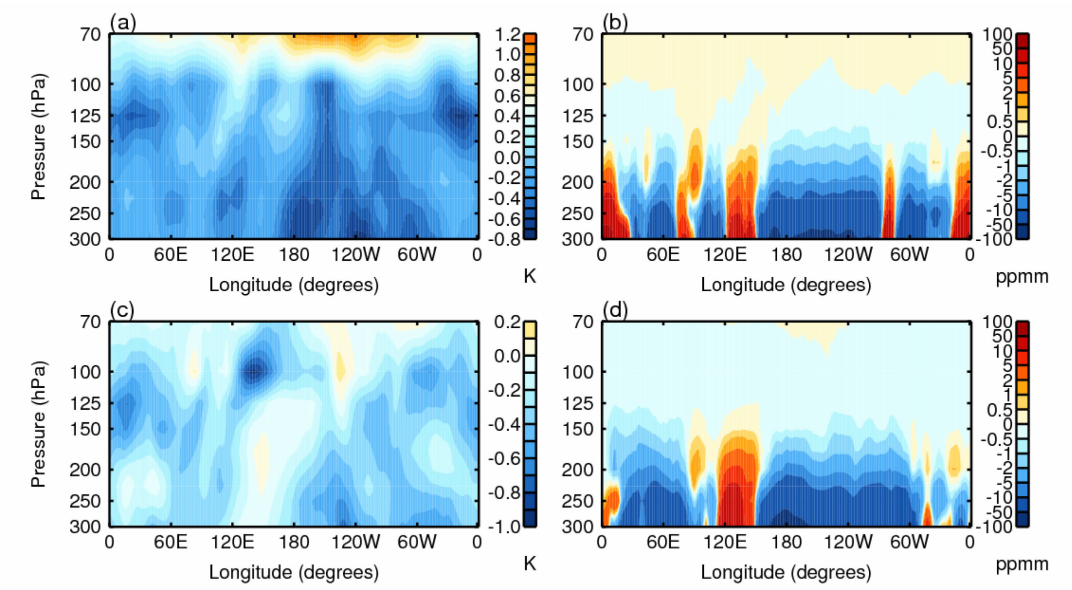

Figure 2. Longitude-height cross-sections of temperature (left) and water vapor (right) anomalies during the (a, b) ‘mismatched’ months and (c, d) ‘matched’ months in 1996.

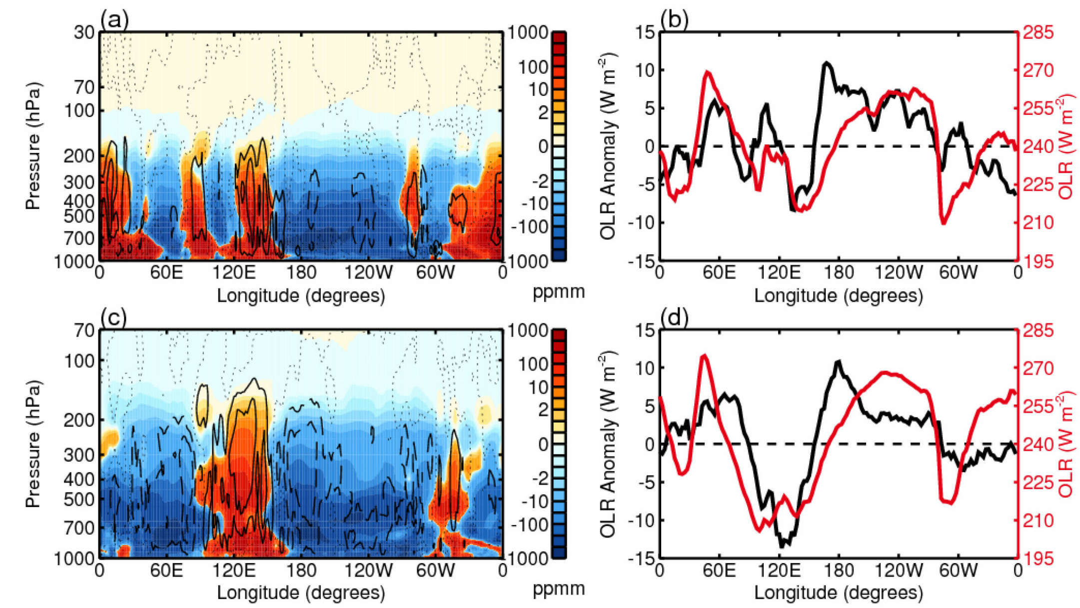

Figure 3. Longitude-height cross-sections of monthly mean water vapor (colors) and vertical velocity (contour levels) anomalies (left), and latitudinal distribution of OLR (red lines) and OLR anomaly (black lines, right) during the (a, b) ‘mismatched’ months and (c, d) ‘matched’ months.

The absolute values of the negative OLR anomaly over the regions 70–100°E and 120–150°E during the ‘mismatched’ months are smaller than those over 100–150°E during the ‘matched’ months, which indicates that the deep convective activity during the ‘matched’ months is stronger. However, under the relatively weak deep convection during the ‘mismatched’ months, water vapor can be transported directly from the lower troposphere into the lower stratosphere (Fig.3a) and is less affected by CPTT, resulting in the poor correlation between CPTT and stratospheric water vapor. Under the relatively strong deep convection during the ‘matched’ months, water vapor can be transported only up to 150 hPa. When water vapor enters the stratosphere in the tropics, a cold tropical tropopause would result in very low water-vapor mixing ratios by effective freeze-drying of the air. Consequently, negative stratospheric water vapor anomalies show a high correlation with the cold tropopause temperature. In summary, during the ‘mismatched’ months, deep convective activity is relatively weak, but water vapor can be transported higher than it is during the ‘matched’ months. The cause of this phenomenon is unknown, and further analysis of this relationship is investigated below.

Previous studies have shown that OLR and highly reflective cloud (HRC) are the indices used most frequently to track deep convective activity in the tropics (Jury [48]; Fu et al.[50]). In general, higher OLR values and lower HRC values correspond to a weakening of deep convective activity, whereas a lower OLR and higher HRC is linked to increased convection (Waliser et al.[51]). Fig.4 shows the time series of daily mean anomalous OLR and HRC averaged over 70–100°E and 120–150°E during the ‘mismatched’ months, and over 100–150°E during the ‘matched’ months. We can see that OLR is strongly correlated with HRC, with a positive (negative) OLR corresponding to a negative (positive) HRC, which is consistent with previous results (Waliser et al.[51]). The correlation coefficients between values of OLR and HRC over 70–100°E, 120–150°E, and 100–150°E are −0.84, −0.80, and −0.81, respectively. A comparison of OLR and HRC values between the ‘mismatched’ (Figs.4a-d) and ‘matched’ months (Figs.4e and f) shows that the occurrence frequency of negative OLR and positive HRC during the ‘mismatched’ months is less than that during the ‘matched’ months. Therefore, the monthly average strength of deep convective activity is weaker during the ‘mismatched’ months. This is consistent with the results shown in Figs.3b and 3d. Close inspection of Figs.4e and 4f shows a higher occurrence frequency of deep convective activity over 100–150°E during the ‘matched’ months, although it is evident from the time-pressure cross-section of vertical velocity averaged over 100–150°E (Fig.5c) that the vertical velocity is not very large during the ‘matched’ months. Under such conditions, water vapor can only be transported up to 150 hPa (Fig.3c). It is worth noting that the instantaneous intensity of the deep convective activity during the ‘mismatched’ months is larger than that during the ‘matched’ months. Specifically, the periods, covered by days 55–57 (March 27–29) and 87–89 (April 28–30) over 70–100°E, and days 22–24 (January 23–25) and 61–63 (April 2–4) over 120–150°E, show instantaneous strong deep convective activity, which is associated with large absolute OLR values and positive HRC anomalies.

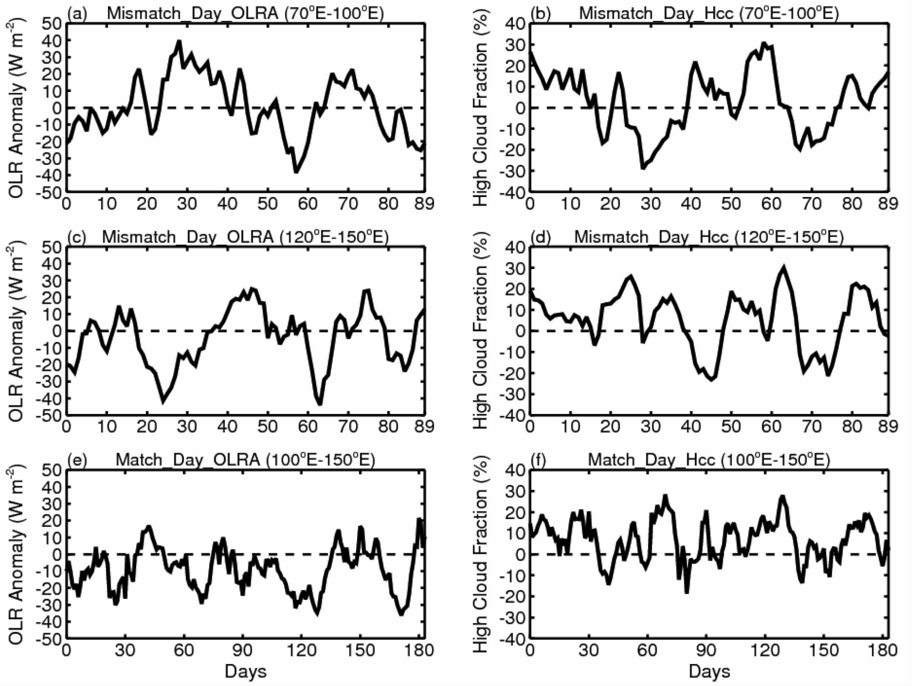

Figure 4. Daily mean time series of OLR anomaly (left) and HRC (right) averaged over (a, b) 70–100°E and (c, d) 120–150°E during ‘mismatched’ months, and over (e, f) 100–150°E during ‘matched’ months.

Figure 5 shows the time–height cross-sections of the daily mean vertical velocity anomaly averaged over 70–100° E and 120–150° E during the ‘mismatched’ months, and averaged over 100–150° E and 60–30° W during the ‘matched’ months. There are positive anomalies in vertical velocity over these four regions, which means the rising motion. This is consistent with the results in Figs.3a and 3c. To be specific, over 70–100° E and 120–150° E during the ‘mismatched’ months, the vertical velocity varies greatly. The strongest vertical velocities of 3.50 and 3.38 m/s occur during days 55–57 and 87–89, respectively, over the 70–100° E region. During the periods covered by days 22–24 and 61–63 over 120–150°E, the strongest vertical velocity can reach up to 4.58 and 2.96 m/s, respectively. This corresponds strongly with the large absolute OLR values and positive HRC anomalies (Figs.4a-d). Nevertheless, during the ‘matched’ months, vertical velocity varies slightly over 100–150°E and most of the vertical velocity centers are between 0.4 and 1.2 m/s, which is in accordance with the results in Figs.4e and 4f. In addition, the vertical velocity over 60–30° W is the smallest over these four regions. Under such conditions, less water vapor is transported, and at lower levels, during the ‘matched’ months than during the ‘mismatched’ months.

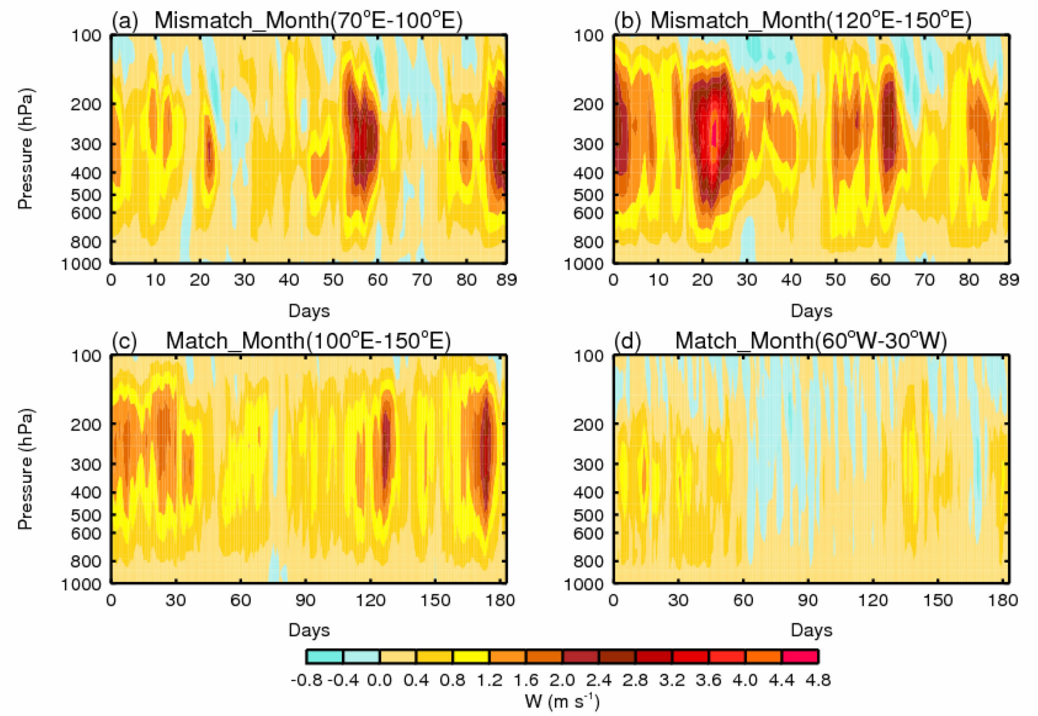

Figure 5. Time-height cross-sections of the daily mean vertical velocity anomaly averaged over (a) 70–100°E and (b) 120–150°E during the ‘mismatched’ months, and averaged over (c) 100–150°E and (d) 60–30°W during the ‘matched’ months.

The above analysis suggests that the occurrence frequency of deep convective activity is less during the ‘mismatched’ months, whereas the instantaneous intensity of the four deep convective centers is greater, and this is seen in the large absolute OLR values and positive HRC anomalies during days 55–57 and 87–89 over 70–100°E, and during days 22–24 and 61–63 over 120–150°E. This indicates that deep convective activity is strong during these four short periods. The enhanced deep convective activity during these four periods can quickly transport lower tropospheric water vapor directly into the lower stratosphere, resulting in an increase in stratospheric water vapor. To confirm this hypothesis, Fig.6 shows the longitude–height cross-section of daily mean water vapor anomalies during days 55–57 and 87–89 over 70–120°E (Figs.6a, b) and during days 22–24 and 61–63 over 120–150°E (Figs.6c and d). Lower troposphere (1,000 hPa) water vapor is transported into the lower stratosphere (70 hPa) during days 87–89 over 70–120°E, and during days 61–63 over 120–150°E, leading to the positive anomaly in stratospheric water vapor. However, there is no evident transport of water vapor in the other two periods; i.e., days 55–57 over 70–120°E and days 87–89 over 120–150° E. As mentioned in the Introduction, the tropical tropopause temperature plays a crucial role in controlling stratospheric water vapor. Combined with the variation in CPTT during these two periods (Figs.7c and d), the absolute values of the negative CPTT anomalies during days 55–57 are larger than those during days 87–89, which indicates a colder tropopause during days 55–57. A colder tropopause suppresses the transport of water vapor from the lower troposphere into the lower stratosphere, leading to there being no evident transport of water vapor. Similarly, a cold tropopause leads to there being no evident transport of water vapor over 120–150°E. With respect to monthly mean water vapor (Fig.3a), there exists transport of water vapor from the lower tropopause into the lower stratosphere over 70–120°E and 120–150°E.

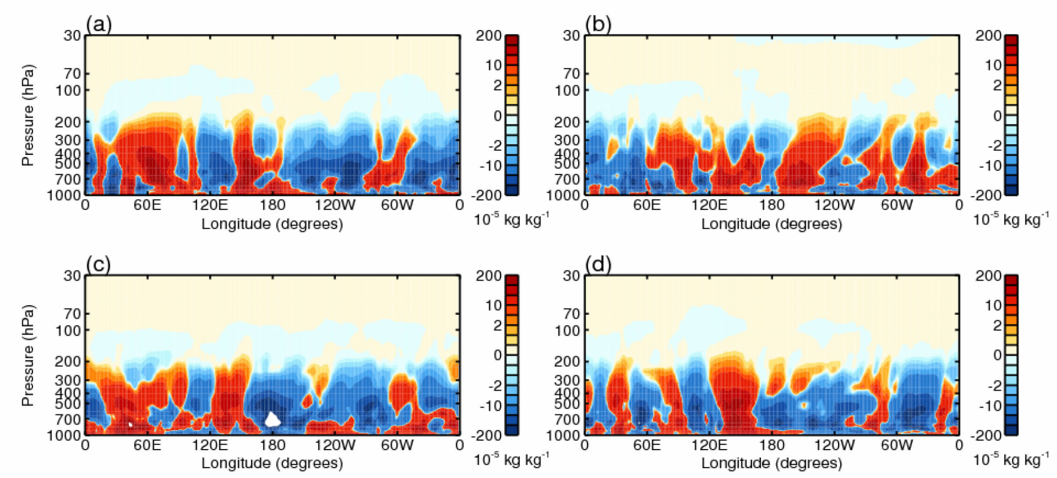

Figure 6. Longitude-height cross-sections of daily mean water vapor anomalies during days (a) 55–57 and (b) 87–89 over 70–120°E, and during days (c) 22–24 and (d) 61–63 over 120–150°E.

Previous studies have shown that deep convective activity can lift the tropopause and cool the tropopause temperature (Gettelman et al.[28]; Teitelbaum et al.[52]). To further diagnose the changes in CPTT during the four instantaneous strong deep convective activity periods, Figs.7a and 7b show the time series of tropical tropopause temperature averaged over 70–120°E and 120–150°E, respectively. Note that the grey areas represent the changes in CPTT during the four periods of instantaneous strong deep convective activity. Negative anomalies in CPTT are evident during these four periods, which implies a lower tropopause temperature. Figs.7c and 7d show latitudinal variations in CPTT averaged over days 55–57 and 87–89, respectively, from Fig.7a. Figs. 7e and 7d are the same as Figs.7c and 7d, but for the means of days 22–24 and 61–63, respectively. The grey areas in Figs.7c–7f represent the regions over 70–120°E and 120–150°E. It is also noticeable that CPTT shows negative anomalies and a declining trend over 70–120°E and 120–150°E, which indicates a decreased CPTT over these two periods. Therefore, Fig.7 suggests that the tropical tropopause temperature decreased during the four periods of instantaneous strong deep convective activity over 70–120°E and 120–150°, which is consistent with previous studies (Gettelman et al.[28]; Teitelbaum et al.[52]).

On the one hand, the instantaneous intensity of the four periods of deep convective activity over 70–120°E and 120–150°E during the 'mismatched' months in 1996 results in the transport of tropospheric water vapor into the lower stratosphere over a short period, which contributes to the increase in stratospheric water vapor. On the other hand, deep convective activity lifts the tropopause and cools the tropopause temperature, and is further responsible for the uncorrelated relationship between tropical tropopause temperature and stratospheric water vapor. In short, the instantaneous intensity of the four periods of deep convective activity is the main factor that influences the uncorrelated relationship in 1996.

To further verify the commonly deep convective activity over 70–120°E and 120–150°E during the 'mismatched' months in 1996, Fig.8 shows the spatial distribution of 200 hPa and 850 hPa wind divergence averaged over the 'mismatched' months in 1996. There is a clear divergence at 200 hPa, but convergence at 850 hPa, over the 70–120°E and 120–150°E regions. Graham and Barnett pointed out the close relationship between deep convective activity and wind divergence [53]. Subsequently, Lau et al. showed that the regions of 200 hPa wind divergence and convergence are closely tied to regions of deep convective activity and to regions of clear sky [54]. Accordingly, upper-level divergence and lower-level convergence favor the occurrence of deep convective activity during the 'mismatched' months in 1996. The spatial distribution of the 850 hPa zonal wind, which is a composite from the negative OLR and positive OLR in Fig.4a, is shown in Fig.9. Strong cyclones occur at 850 hPa over 70–120°E when compositing the negative OLR (box in Fig. 9a), whereas there is an absence of cyclones when compositing the positive OLR (box in Fig.9b). The strong cyclones would induce the convergence and rising motion, and so contribute to the instantaneous intensity of the four periods of deep convective activity.

Figure 10 further shows the distribution of SST and

OLR, and areas of high SST coincide with regions where OLR is decreased and deep convective activity is enhanced, and vice-versa for the low SST. Earlier studies indicated a close relationship between SST and deep convective activity (Gutzler and Wood [55]; Zhang [56]). Zhang pointed out that the frequency and strength of deep convective activity increases substantially with changes in SST from 26.5℃ up to 29.5–30℃[56]. Lau et al. further confirmed that OLR decreases significantly as SST increases, when SST varies between 26.5℃ and 29.5℃[54]. The mean SST over 10.5°N–10.5°S and 120–150° E during the 'mismatched' months is 28.6℃ , Hence, a high SST may be the main factor which causes the enhanced deep convective activity over 120–150°E.

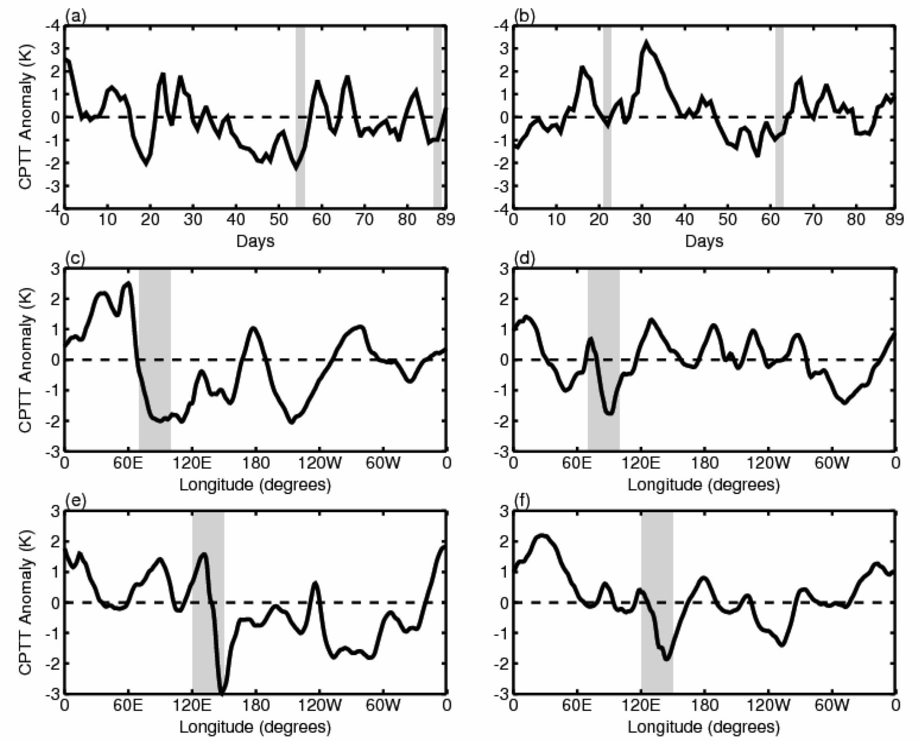

Figure 7. (a, b) Time series of tropical tropopause temperature averaged over (a) 70–120°E and (b) 120–150°E. Grey areas represent changes of CPTT during the four periods of instantaneous strong deep convective activity. (c-f) Latitudinal distribution of CPTT during days (c) 55–57 and (d) 87–89 over 70–120°E, and during days (e) 22–24 and (f) 61–63 over 120-150° E.

We also investigated the spatial distribution of annual-mean zonal wind and SST during the rest of the 'matched' years between 1994 and 2005 (not shown here). The zonal wind and SST are calculated in the same way as for Figs.9 and 10. We find that strong surface cyclones and high SST do not develop over 70–120°E and 120–150°E during the rest of the 'matched' years. This confirms that the instantaneous intensity of the four periods of deep convective activity, caused by strong surface cyclones and high SST, is the main factor that results in the poor correlations between tropical tropopause temperature and stratospheric water vapor.

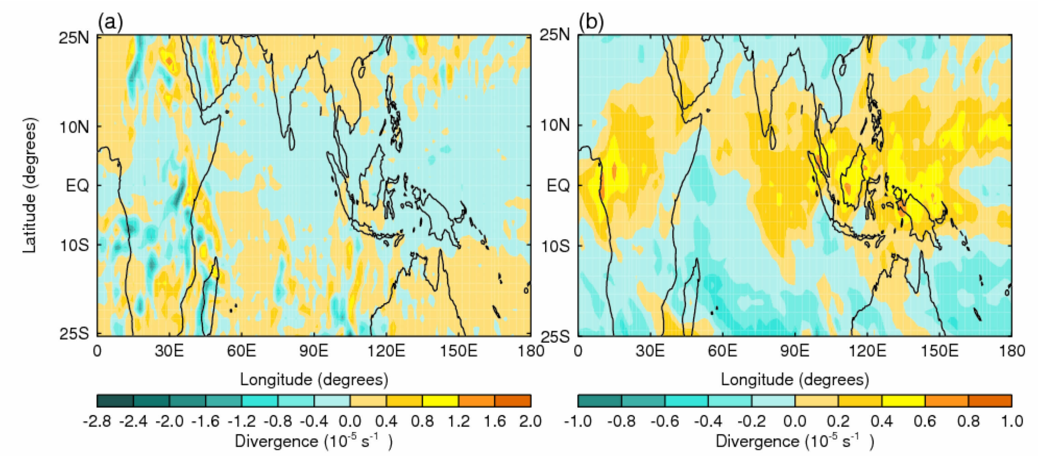

Figure 8. Spatial distribution of (a) 200 hPa and (b) 850 hPa wind divergence during the 'mismatched' months in 1996.

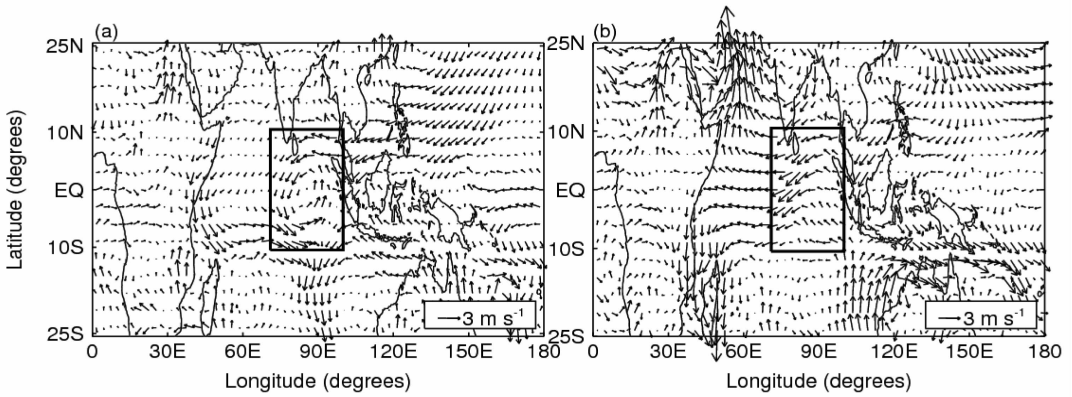

Figure 9. Spatial distribution of daily mean 850 hPa zonal wind obtained by compositing the (a) negative OLR and (b) positive OLR during the ‘mismatched’ months in 1996. The box represents the area over 10.5°N–10.5°S and 70–100°E.

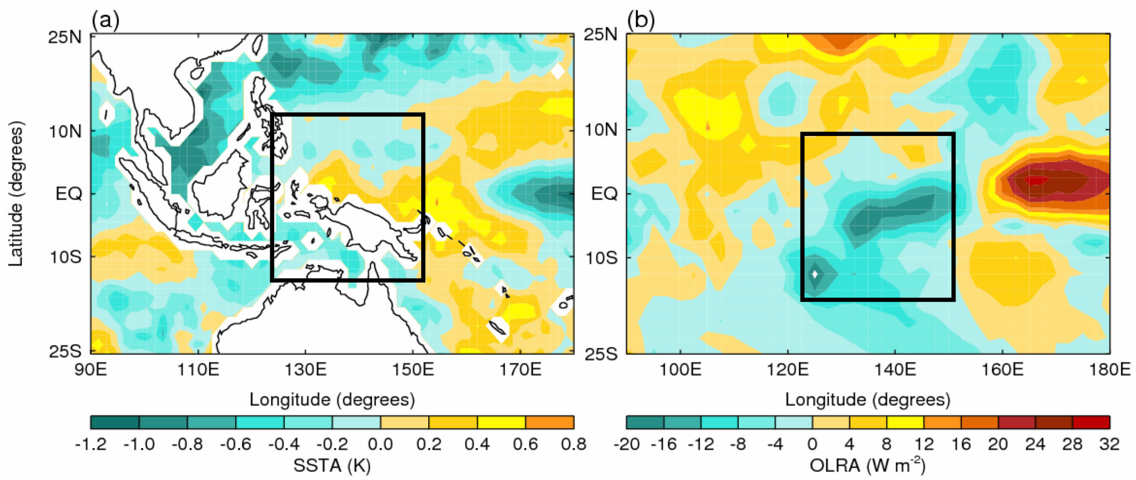

Figure 10. Spatial distribution of the monthly mean anomaly of (a) SST and (b) OLR during the ‘mismatched’ months in 1996. The box represents the area over 10.5°N–10.5°S and 120–150°E.

In addition, BDC is a dominant control on water vapor transport in the stratosphere (Chen and Chen[57]; Dhomse et al.[58]). Fig.11 shows the vertical component of the BDC ($\overline{\omega}^*$, left column) and latitude-height cross-sections of the vertical component of water vapor flux ($\overline{\omega}^* q$, right column) during the ‘mismatched’ months and the ‘matched’ months in 1996. During the ‘mismatched’ months, a large positive anomaly of $\overline{\omega}^*$ occurs almost everywhere in the tropics from 200 to 10 hPa, except between 5°S and 10.5°S. This indicates that the BDC is enhanced during the ‘mismatched’ months in 1996, which indicates a strengthening tropical upwelling. In contrast, $\overline{\omega}^*$ during the ‘matched’ months is less than that during the ‘mismatched’ months, suggesting a weakening tropical upwelling.

4 CONCLUSIONS

In this study, we use measurements from HALOE and the ERA-Interim reanalysis data for the period 1994–2005 to analyze the relationship between tropical tropopause temperature anomalies and stratospheric water vapor anomalies. Further analyses are also completed into the cause of the uncorrelated relationship between tropical tropopause temperature and stratospheric water vapor. Our major findings can be summarized as follows.

(1) The tropical tropopause temperature is correlated with stratospheric water vapor during the period 1994–2005, except for 1996 and 1999.

(2) The instantaneous intensity of the four periods of deep convective activity in February, March, and April 1996, which are caused by strong surface cyclones and high SST, results in two water vapor transport channels over 70–120°E and 120–150°E. Water vapor is transported from the lower troposphere to the lower stratosphere by the strong deep convective activity, leading to an increase in stratospheric water vapor. Meanwhile, deep convective activity lifts the tropopause and cools the tropopause temperature, leading to their uncorrelated relationship. In contrast, during the ‘matched’ months, a transport channel for water vapor occurs at around 100–150°E, but water vapor can only

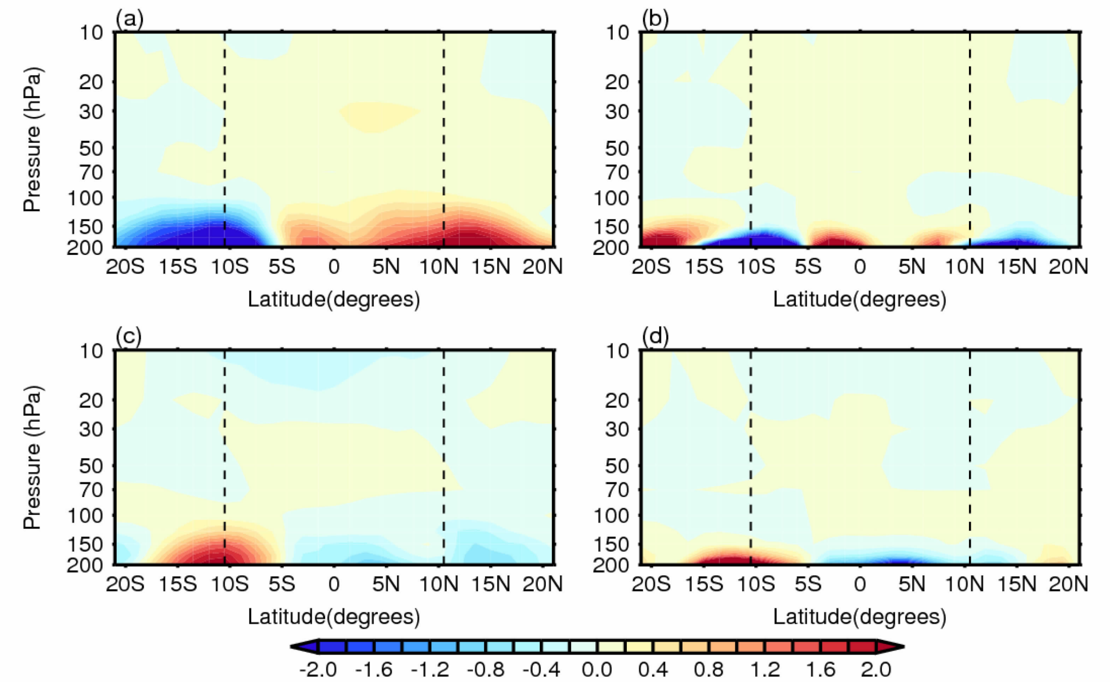

Figure 11. Latitude-height cross-sections of monthly mean anomalies of (left) and (right) during (a, b) the 'mismatched' months and (c, d) the 'matched' months in 1996. Areas between two black dotted lines represent the tropics (10.5°N–10.5°S).

be transported up to 150 hPa. When water vapor enters the stratosphere in the tropics, a cold tropical tropopause would result in very low water-vapor mixing ratios due to the effective freeze-drying of the air; thus, negative anomalies of stratospheric water vapor show a high correlation with a cold tropopause temperature. The increase in water vapor during the 'mismatched' months is larger than the decrease in water vapor during the 'matched' months, indicating an increase in annual-mean water vapor. Therefore, there is an uncorrelated relationship between tropical tropopause temperature and stratospheric water vapor in 1996.

(3) Comparison of the 'mismatched' months and the 'matched' months shows that the occurrence frequency and strength of the deep convective activity during the 'mismatched' months are less and weaker than that during the 'matched' months in 1996. However, the instantaneous intensity of the four periods of deep convective activity is greater during the 'mismatched' months than during the 'matched' months. Water vapor is transported from the lower troposphere to the lower stratosphere through four periods of instantaneous strong and deep convective activity, leading to an increase in stratospheric water vapor. In contrast, water vapor is only transported up to 150 hPa during the 'matched' months.

(4) An anomalously strong Brewer-Dobson circulation brings more water vapor into the stratosphere during the 'mismatched' months in 1996, further leading to the poor correlation between tropical tropopause temperature and stratospheric water vapor.

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