Temperature Fields in the Tropical Tropopause Transition Layer

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(Manuscript received 20 June 2003, in final form 29 January 2004)

ABSTRACT

There has been increasing recognition of the role of the tropical tropopause layer (TTL) in determining stratospheric water vapor concentrations—the TTL being a layer of transition between air showing tropospheric properties below and stratospheric properties above. This study investigates the spatial structure of temperatures in the TTL. A dehydration index based on the atmospheric region with temperatures colder than a specific reference temperature was defined to examine the TTL temperature structure and possible influences on stratospheric water vapor. The results indicate that dehydration regions with cold temperatures (e.g., $<$190 K) occur mainly over the western Pacific and are about 1.5–2.0 km in depth during Northern Hemisphere winter. The dehydration index is mainly dependent on the annual cycle of the TTL temperatures, but is strongly affected by interannual variations associated with the quasi-biennial oscillation (QBO) and El Niño–Southern Oscillation (ENSO). Dehydration regions with extremely cold temperatures and large sizes occur when cold temperature anomalies associated with the QBO arrive at the TTL in wintertime while the TTL is at the coldest phase of the annual cycle and under La Niña conditions. La Niña events have a more dramatic influence on the dehydration index than El Niño events.

1. Introduction

Mass exchange between the troposphere and the stratosphere is dominated by the mean meridional circulation, with upwelling in the Tropics and downward motion at higher latitudes (Holton et al. 1995, and references therein). The fact that the observed concentration of stratospheric water vapor is much smaller than the saturation mixing ratio of water vapor for zonally averaged tropical temperatures led Newell and Gould-Stewart (1981) to propose the ''stratospheric fountain'' hypothesis. This suggests that tropospheric air enters the stratosphere across the tropical tropopause over the western Pacific, and preferentially during the Northern Hemisphere (NH) winter. However, some studies indicated that the vertical velocity near the tropopause over the western Pacific is likely downward instead of upward (Sherwood 2000). Thus, it is more appropriate to call the cold tropical tropopause layer (TTL) area over the western Pacific the ''cold trap'' for air entering the stratosphere (Holton and Gettelman 2001). In this paper, the term ''dehydration regions'' was used instead of the term cold trap when referring to cold TTL areas in tropical longitudes including, but not limited to, the western Pacific.

During the past few years, there has been increasing recognition of the role of the TTL in determining stratospheric water vapor—the TTL being a layer of transition between air showing tropospheric properties below and stratospheric properties above (Sherwood and Dessler 2001; Holton and Gettelman 2001). In this study, we investigate the spatial and temporal structure of temperatures near the tropical tropopause. Since we are considering dehydration to stratospheric water vapor concentrations, we defined a ''dehydration volume'' index (DV) based on the atmospheric region with temperatures colder than a specific reference temperature to examine the TTL temperature structure and its possible influence on stratospheric water vapor. This index reflects the degree of coldness of the TTL. We found that DV during the 1982/83 winter was very small and DV in the 1984/85 winter was very large. Previous studies indicate that the tropical tropopause is affected by the stratospheric quasi-biennial oscillation (QBO) and the tropospheric El Niño–Southern Oscillation (ENSO; Randel et al. 2000; Zhou 2000; Zhou et al. 2001b). In this study, we investigate the combined effects of the QBO and ENSO, as manifested in the temperature field near the tropical tropopause.

This paper is arranged as follows: section 2 briefly describes the definition of DV and the datasets used in this study, and section 3 presents horizontal and vertical distributions of multiyear mean temperature fields near the tropopause. The time series of DV is given in section 4. Also in this section, the underlying physical circumstances for two extreme cases are presented. Section 5 is the discussion and summary section.

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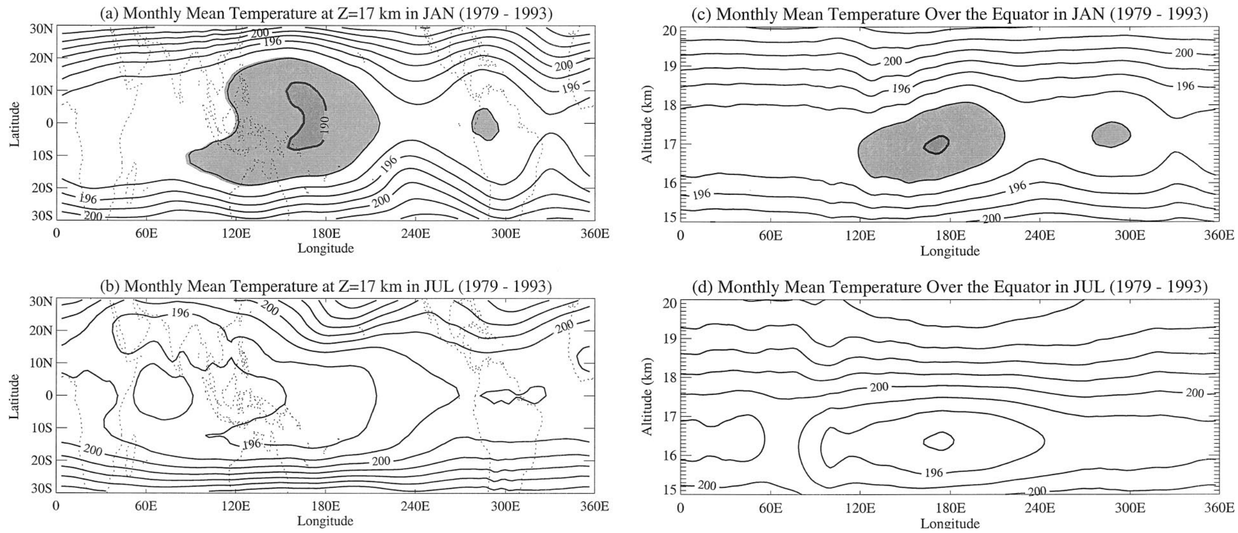

FIG. 1. Sections of temperatures near the tropical tropopause in Jan and Jul averaged over the ERA-15 available period (1979–93): (a) horizontal section at 17 km in Jan; (b) similar to (a), but for Jul; (c) vertical section over the equator in Jan; (d) similar to (c), but for Jul. Areas with temperature below 192 K are shaded.

2. Datasets and definition of the dehydration volume

The principal dataset used in this study is the 15-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-15: 1979–93). Stratospheric zonal wind shear at 50 hPa over Singapore and the sea surface temperature anomaly (SSTA) in the Niño-3.4 region (5°S–5°N, 120°–170°W) were used as indices for the QBO and ENSO, respectively.

To investigate variations in the coldness of the TTL, a dehydration volume index DV was defined as the volume of air with temperatures below the reference temperature in the tropical region between 15–20 km and 20°S–20°N. The grid version of ERA-15 has 17 standard pressure surfaces and a horizontal resolution of 2.5° $\times$ 2.5°. Temperatures every 100 m in the vertical direction between 15–20 km and 20°S–20°N were obtained using cubic spline fitting (refer to Zhou et al. 2001a for more information on this). Comparison with sounding data indicated that ERA-15 tracked tropopause variations very well except that it overestimated the tropical cold-point tropopause temperatures by approximately 2 K (Zhou et al. 2001b). The reference temperature adopted is 192 K. Considering the warm bias in ERA-15, we strived for an actual reference temperature of approximately 190 K. It should be emphasized that the warm bias of temperature in ERA-15 was not adjusted in the temperature sections that are shown in this paper. Assuming a pressure of 100 hPa, the real reference temperature (190 K) corresponds to a saturation mixing ratio of water vapor with respect to ice of approximately 3.26 ppmv. The entry value of water vapor mixing ratio at the tropical tropopause is about 3.0–4.1 ppmv based on multiple instrument observations, corresponding to a 100-hPa entry-level temperature range of 189.7–191.4 K (Kley et al. 2000). The reference temperature used in DV calculations is close to the lower value of this temperature range.

3. Mean temperature field near the tropical tropopause

Figure 1 shows average horizontal sections of ERA-15-based temperatures at 17 km and vertical sections of temperatures over the equator. The areas with temperatures below the reference temperature (192 K) are shaded. It is seen that cold temperatures occur over the western Pacific in NH winter. In NH summer, tropopause temperatures are about 4 K warmer, and the tropopause is about 0.5 km lower than in winter. This is consistent with previous results (e.g., Newell and Gould-Stewart 1981). Figure 1 suggests that the cold trap is not only extensive horizontally but is also thick (about 1.5–2.0 km). Upwelling velocities near the tropical tropopause are about 0.1–0.2 mm s^{-1} in July and 0.2–0.4 mm s^{-1} in January (Rosenlof 1995; Mote et al. 1998). Thus, it takes about 3–6 months for an air parcel to transverse a depth of 1.5 km near the tropopause at the typical summertime upwelling velocity and about 1–3 months at the typical wintertime upwelling velocity. It also takes about 1–2 months for an air parcel to travel once around the earth along the equator, assuming a zonal wind of 5–10 m s^{-1}. Because of the fast horizontal motion and slow vertical motion, air parcels that cross the lower boundary of the TTL at longitudes other than those in the western Pacific have a large probability of being dehydrated by the cold trap before they enter the stratosphere across the upper boundary of the TTL. Because summertime upwelling velocities are much slower than wintertime upwelling velocities, many air parcels that

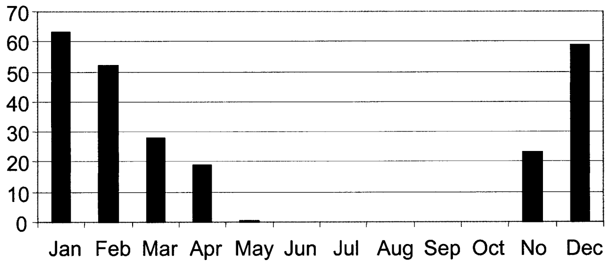

FIG. 2. The annual cycle of DV (10^6 km^3) based on ERA-15. The DV is defined as the volume of air with temperatures below a reference temperature in the tropical region between 15–20 km and 20°S–20°N.

cross the lower boundary of the TTL in summer are dehydrated by the cold trap in winter before they enter the stratosphere. Of course, not all summertime air parcels are dehydrated by the wintertime cold trap. Otherwise, the ''tape recorder'' in stratospheric water vapor would not have been observed (Mote et al. 1996). While the above speculation is quasi quantitative, trajectory calculations using accurate upwelling velocities and horizontal winds are needed to rigorously show that it is true.

Values for DV, based on monthly mean temperatures averaged over the ERA-15 period (1979–93), were plotted in Fig. 2. It can be seen that large DV values occur during NH winter, with the largest value of 63.07×10^6 km^3 occurring in January. During NH summer, the average DV for these years is zero due to the relatively warm tropopause layer temperatures at that time.

4. Variations of DV and two extreme cases

Figure 3 shows the time series of DV derived from ERA-15. Values of DV for the western Pacific (120°–210°E) are also plotted. The DV has large values during the NH winter seasons and is either virtually or identically zero during the NH summer seasons. This is consistent with previous results indicating that the tropical tropopause is much warmer during NH summer than in winter. For most winters, DV for the entire Tropics was only slightly larger than DV for the western Pacific. This is consistent with previous results indicating that the coldest tropopause regions are over the western Pacific during NH winter (Newell and Gould-Stewart 1981). However, there were large differences between the DV values for the entire Tropics and the western Pacific for some winters, especially the 1984/85 winter. The value of DV for the entire Tropics was about 280 $\times 10^6$ km^3 and for the western Pacific was about 100 $\times 10^6$ km^3 during that winter. Both values are much larger than the multiyear mean DV value in January (63.07×10^6 km^3). The large difference between the DV values for the entire Tropics and the western Pacific indicates that, during the 1984/85 winter, the dehydration regions had a broad longitudinal extension beyond the western Pacific. For another extreme case, DV in 1982/83 was very small for both the entire Tropics and the western Pacific. The ERA-15-based DV calculation is consistent with sounding observations. The tropical cold point tropopause (CPT) temperature averaged over the entire Tropics using operational sounding observations indicated that the 1984/85 winter was one of coldest winters, and that the 1982/83 winter was one of warmest winters during the 1973–98 period (see Fig. 3 of Zhou et al. 2001a).

Figure 4 shows the horizontal sections at 17.5 km and vertical sections over the equator for the two winters, 1982/83 and 1984/85. The 17.5-km level was selected because it is close to the average altitude of the tropical CPT in NH winter. During the 1982/83 winter, the cold area with temperatures below the reference temperature (192 K) covered a small area over the subtropical eastern Pacific to the tropical Atlantic (Fig. 4a) and was relatively thin in its vertical extent (Fig. 4b). During the

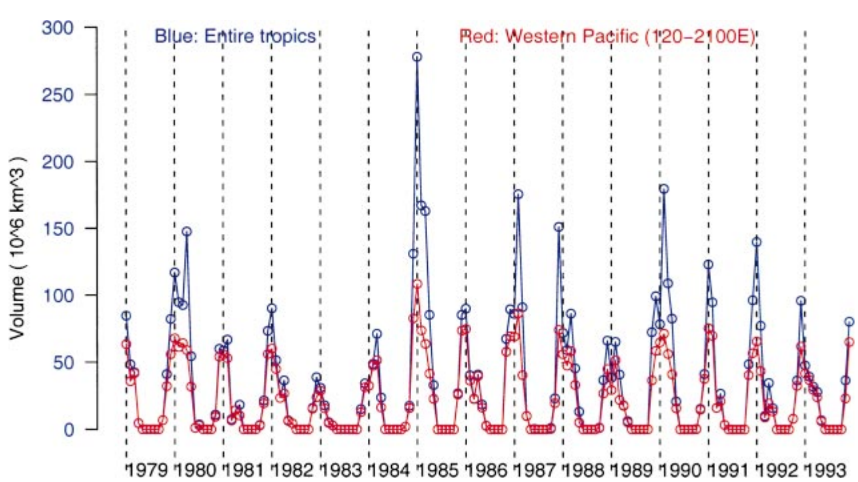

FIG. 3. Dehydration volume based on ERA-15 monthly mean temperatures. The blue curve is DV for the entire Tropics and the red curve is for the western Pacific (120°–210°E).

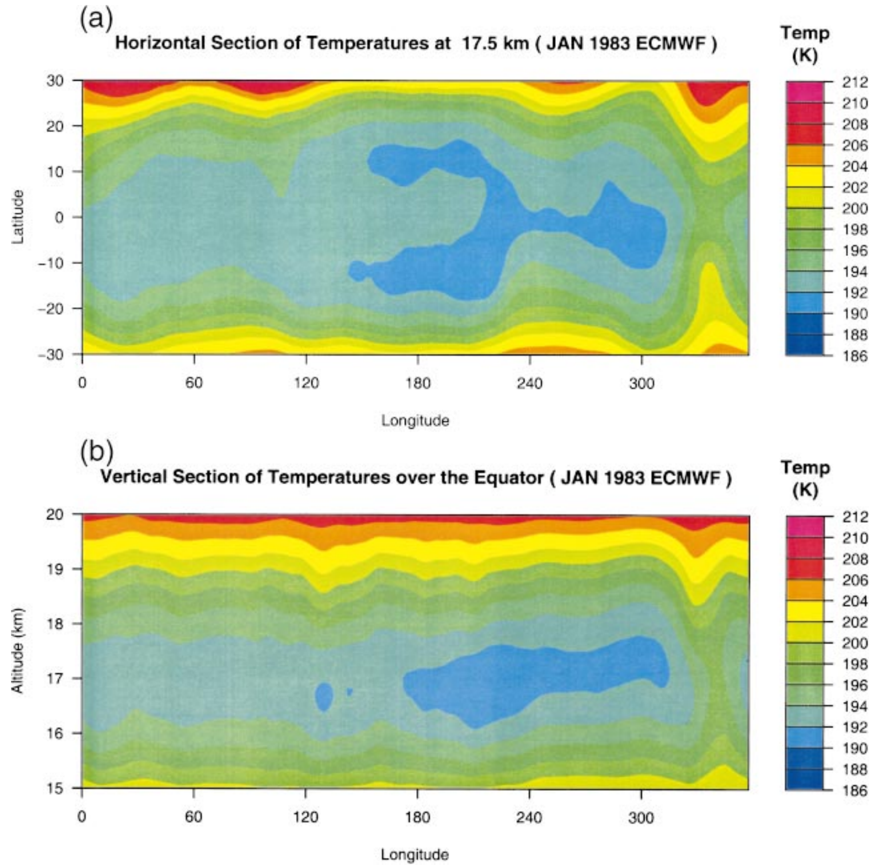

FIG. 4. Horizontal sections at 17.5 km and vertical sections over the equator of the TTL temperatures in two extreme cases: (a) horizontal section for Jan 1983, (b) vertical section for Jan 1983, (c) horizontal section for Jan 1985, and (d) vertical section for Jan 1985.

1984/85 winter, the TTL was cold throughout the entire Tropics, with the coldest area below 188 K (or below 186 K after the warm bias is allowed for) near the date line (Fig. 4c). The cold area was relatively thicker (Fig. 4d), and the level of the CPT was about 0.5 km higher than in 1982/83. The Stratospheric Processes and their Role in Climate (SPARC) Water Vapor Assessment indicated that the entry value of water vapor at the tropical tropopause is about 3.0–4.1 ppmv, corresponding to a 100-hPa entry-level temperature range of 189.7–191.4 K (Kley et al. 2000). Monthly mean CPT temperatures at almost any longitude in the 1984/85 winter were able to dehydrate air parcels entering the stratosphere to the lower range of the estimated entry value of water vapor.

Figure 5a shows the differences of temperature between the two winters at 17.5 km. Temperature differences greater than 4 K occurred at most longitudes. Note that an El Niño occurred in 1982/83 and a La Niña occurred in 1984/85. The horizontal section of the temperature differences is dominated by the ENSO signature in temperatures near the tropopause (Zhou et al. 2001b). The ENSO signature at the tropical CPT shows features of a north–south dumbbell, a west–east dipole, and a cold–warm core between the dipole (Zhou et al. 2001b). During El Niño periods, the tropopause has cold anomalies over the eastern Pacific and warm anomalies over the western Pacific. However, the tropopause over the eastern Pacific is usually 4 K warmer than the tropopause over the western Pacific (Fig. 1a). This horizontal contrast of temperatures near the tropopause is larger than the magnitude of the cold temperature anomalies associated with El Niño over the eastern Pacific. Therefore, the cold TTL temperature anomalies associated with El Niño events over the eastern Pacific are not able to increase DV. The warm tropopause temperature anomalies associated with El Niño over the western Pacific warm this climatologically cold area. As a result, El Niño generally results in a smaller DV value. In contrast, during La Niña events, the warm TTL temperature anomalies over the eastern Pacific do not significantly affect DV because, climatologically, this area is already warmer than the defined temperature threshold. Cold temperature anomalies over the western Pacific (the cold ''dumbbell'' in the subtropics and the cold core between the east–west ''dipole'' along the equator; see Fig. 8b of Zhou 2001b) make this climatologically cold region even colder. This results in the cold trap covering a wider area and being thicker in the

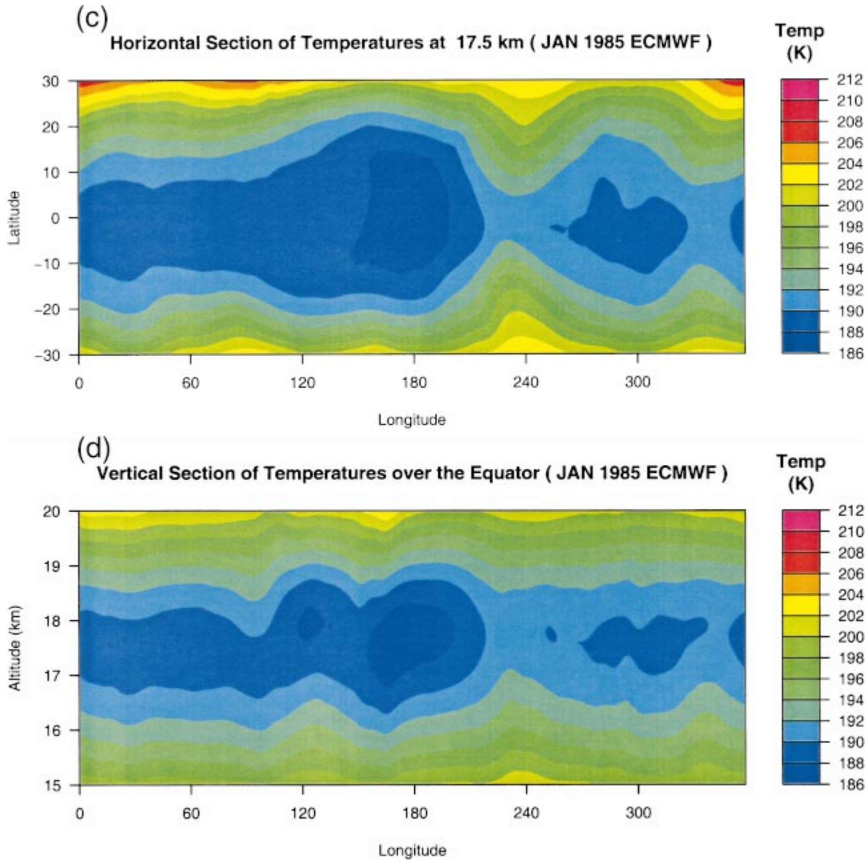

Fig. 4. (*Continued*)

vertical direction. As a result, DV increases during La Niña events. Thus, positive (negative) SSTAS in the Niño-3.4 region generally represent negative (positive) contributions of ENSO events to DV.

The vertical section of the temperature differences shows cold and warm downward-propagating features along the equator (Fig. 5b). Figure 6 shows the temporal evolution of temperature anomalies at altitudes between 10 and 25 km over the equator. The anomaly here was defined as the zonal monthly temperature at each altitude minus its annual cycle. The dominant feature of Fig. 6 is alternating positive and negative temperature anomalies propagating downward from the stratosphere to the TTL. QBO warm temperature anomalies are associated with westerly wind shears, and QBO cold anomalies are associated with easterly wind shears (Plumb and Bell 1982). It takes about 4–7 months for the wind shear or QBO temperature anomaly to reach the tropopause. Accordingly, a 6-month time lag adjustment for the QBO wind shear time series was made in Fig. 7, which shows the time series for normalized SSTAs in the Niño-3.4 region and stratospheric wind shear at 50 hPa. The cold area (e.g., $T < 192$ K) is located in the altitude range of 16–18 km (Fig. 1c) rather than at a fixed altitude. Also, westerly wind shear propagates downward faster than the easterly shear (Naujokat 1986). Thus, the 6-month adjustment for time lag is only a rough estimate. However, this does not affect the following interpretation in a significant way.

The QBO wind shear in Fig. 7 is hatched with horizontal lines and the SSTA index is hatched with vertical lines. The period with positive (negative) contributions of either the QBO wind shear or the SSTA to the tropopause temperatures (DV) is indicated in red, and negative (positive) contributions to the tropopause temperatures (DV) are indicated in blue. A westerly wind shear coupled with a warm Niño-3.4 SSTA anomaly (i.e., El Niño) was present during the 1982/83 winter. Both factors tend to reduce DV. An easterly wind shear coupled with a cold Niño-3.4 SSTA anomaly (i.e., La Niña) occurred during the 1984/85 winter. Both factors tend to increase DV, resulting in extremely large DV values. Thus, the extremely large DV in the 1984/85 winter and the extremely small DV in the 1982/83 winter were caused by temperature anomalies associated with both QBO and ENSO. Similar coupling between QBO and ENSO implied that the 1996/97 (1997/98) winter was another colder (warmer) winter for the tropical tropopause (Fig. 7). Sounding observations showed that the tropical CPT in the 1996/97 (1997/98) winter was indeed colder (warmer) than normal, comparable to the

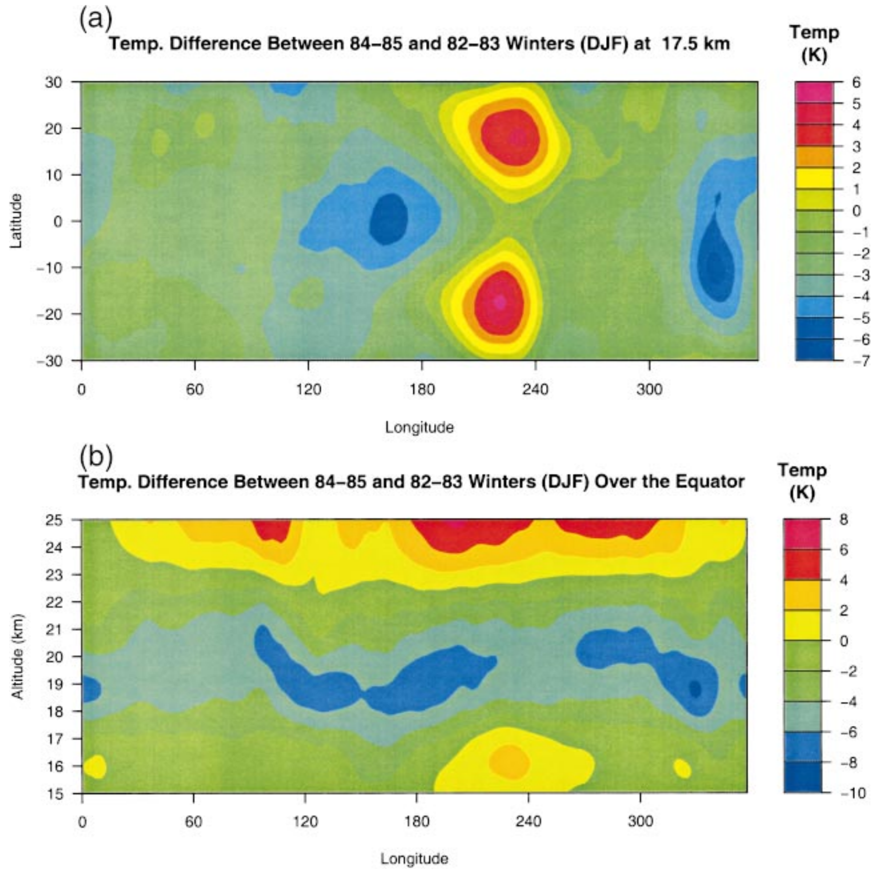

FIG. 5. (a) Horizontal and (b) vertical sections of the temperature difference between the 1984/85 and 1982/83 winters (Dec–Jan–Feb, and the 1984/85 winter minus the 1982/83 winter).

1984/85 and 1982/83 winters, respectively (Fig. 2 of Zhou et al. 2001b; Fig. 3 of Zhou et al. 2001a).

We also calculated DV using the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (NCEP-R: 1979–2002). The NCEP-R-based DV was consistent with the ERA-15-based DV during the available period of the ERA-15 (figure not shown). The NCEP-R-based DV indicates a change of the DV pattern after the early 1990s. The wintertime DV increased, and there were noticeable DV values in summertime, whereas these were absent before this time. The cause for this remarkable change in DV, whether it is a spurious feature in the reanalysis or a true physical variation (Christy et al. 2003), requires further investigation.

5. Discussion and summary

In this paper, the temperature field near the tropical tropopause and the volume of extremely cold regions were studied. Results indicated that temperatures near the tropical tropopause are much warmer during NH summer than in winter. According to sounding observations, cold tropopause temperatures may fall below the reference temperature (190 K) in the summer; however, they occur much less frequently than in winter. Based on monthly mean data, summertime DV is virtually zero for a stratospheric dehydration reference temperature (e.g., 190 K, or 192 K before the ERA-15 warm-bias adjustment was made). In the winter, the cold trap is over the western Pacific and has a thickness of about 1.5–2.0 km in the vertical. Given the small upwelling velocity of the meridional circulation and fast zonal motion near the tropical tropopause, air parcels that enter the TTL across its lower boundary at longitudes other than the western Pacific are likely dehydrated by the cold trap over the western Pacific before they enter the stratosphere across the upper boundary of the TTL. Also, many of the air parcels that enter the TTL in the summer are dehydrated by the cold trap in winter. The latter suggests that wintertime tropopause temperatures are more important than summertime tropopause temperatures for dehydration to stratospheric water vapor mixing ratios. This needs to be further explained with detailed trajectory calculations.

To first order, the QBO signature in the tropopause temperatures is zonally symmetric; however, the ENSO signature is zonally asymmetric (Randel et al. 2000; Zhou et al. 2001b). It has a north–south dumbbell, west–east dipole pattern, and an area of maximum anomalies

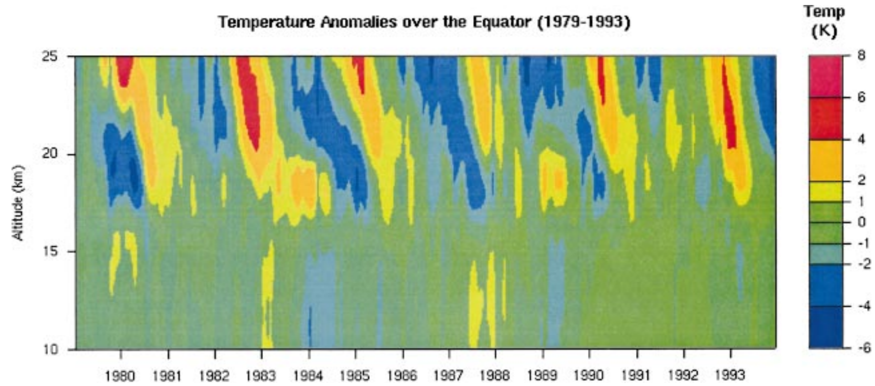

FIG. 6. Evolution of temperature anomalies over the equator. The anomalies are defined as zonal mean temperatures minus their annual cycles at each altitude. The tick marks along the horizontal axis correspond to Jan of each year.

between the west–east dipole (Zhou et al. 2001b). If the cold anomalies associated with the QBO arrive at the tropopause in summer, the QBO will not make a significant contribution to DV because the background tropopause temperatures (i.e., mean temperatures in the summer) are very warm. The summer of 1992 is an example of this. If the cold anomalies associated with the QBO reach tropopause levels in winter, when the annual cycle of tropopause temperatures is at its coldest phase, the coupling between cold QBO anomalies and the annual cycle of the tropopause temperatures enhances DV. On the other hand, if warmer QBO anomalies reach the TTL in winter, it will reduce DV values at that time.

The contribution of ENSO events to DV variations can be examined by looking at the degree of zonal asymmetry of tropopause temperatures. The eastern Pacific is normally about 4 K warmer than the western Pacific in winter. When winter occurs during an El Niño period, the El Niño makes the eastern Pacific a little colder and the western Pacific a little warmer, reducing the zonal temperature contrast. This leads to a low DV. On the other hand, when winter occurs during a La Niña period, the maximum negative equatorial anomalies over the western to central Pacific (around 150°–160°E) and the associated cold dumbbell-shaped anomalies (around 100°–120°E) make the tropopause over the subtropical western Pacific colder, resulting in a larger area of the cold trap. Thus, La Niña events provide a greater and colder region over which dehydration can occur than exists during El Niño events.

Two extreme cases were investigated in this paper. The small DV in the 1982/83 winter was caused by the arrival of both warm QBO and El Niño temperature anomalies at CPT altitudes. The extremely large DV in the 1984/85 winter was due to the additive effects of the cold QBO phase and La Niña–related temperature anomalies with the cold phase of the annual temperature cycle. Large DV and small DV values were expected during the 1996/97 and 1997/98 winters, respectively. Sounding observations show some supportive evidence for this. However, ERA-15 does not cover these years.

Heavy aerosol loading near the tropical tropopause makes the tropopause warmer and leads to smaller DV values (e.g., Angell 1993; Randel et al. 2000). The tropical tropopause became warmer after the El Chichón

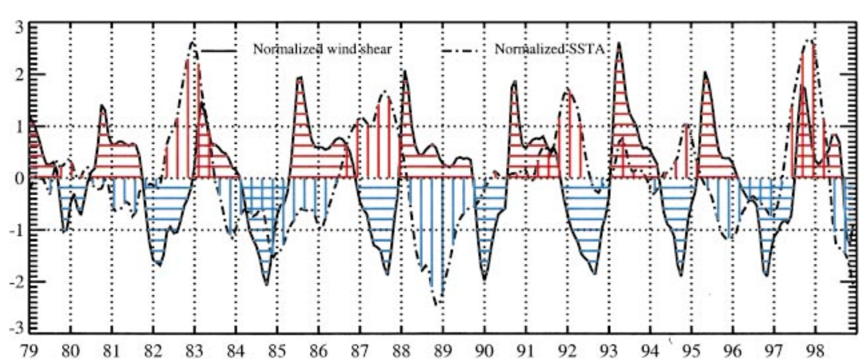

FIG. 7. Normalized time series of stratospheric wind shear at 50 hPa over Singapore and SSTAs in the Niño-3.4 region. Areas enclosed by the wind shear curve are hatched with horizontal solid lines and areas enclosed by the SSTA curve are hatched by vertical solid lines. Red was used for positive contributions to DV and blue was used for negative contributions.

volcanic eruption in 1982; however, the tropical tropopause did not show large warm anomalies after the Mt. Pinatubo eruption in 1991 (Fig. 3 of Zhou et al. 2001a). This difference might be caused by different phases of the QBO. After the El Chichón eruption, warm temperature anomalies associated with the QBO arrived at the tropopause levels and enhanced the warning at the tropical tropopause. However, cold temperature anomalies associated with the QBO arrived at the tropopause after the eruption of Mt. Pinatubo, offsetting the warming aerosol effect. The observed cooling trend of tropical tropopause temperatures also offset the warming effect of the eruption of Mt. Pinatubo. The El Chichón eruption contributed to the extremely small DV value in the 1982/83 winter, but it is not known how important the volcanic eruption is to DV compared with the QBO influence. This is worth further investigation; however, it is beyond the scope of this paper.

Multiple processes affect the tropical tropopause, including its annual cycle, the QBO, ENSO, volcanic eruptions, trends, etc. The potential contribution of any of these processes might be studied by neglecting the other processes or keeping them unchanged (e.g., Geller et al. 2002; Joshi and Shine 2003). Due to the strong nonlinearity of the Clausius–Clapeyron equation, however, a small anomaly in tropical tropopause temperatures, regardless of what process causes it, results in a significant change in the saturation water vapor mixing ratio. The net anomaly of contributions from multiple sources does not necessarily behave in the same way as any of the processes acting alone due to the enhancement and/or cancellation among the processes. The entry value of water vapor mixing ratio across the tropical tropopause (and thus the trend of stratospheric water vapor) depends on the temperature annual cycle and the net anomaly. The two extreme cases investigated in this paper show that coupling between QBO and ENSO during NH winters (when the tropical tropopause is in the cold phase of its annual cycle) may change temperature fields near the tropical tropopause dramatically and result in extremely large or small DV values. These extremely large or small DV values should have been ''recorded'' in stratospheric water vapor. This needs further investigation.

In summary, the cold trap over the western Pacific has a thickness of about 1.5–2.0 km in NH winter. Wintertime tropopause temperatures are more important than summertime tropopause temperatures for stratospheric dehydration, given slow tropical upwelling. An index in determining the yearly averaged entry water vapor mixing ratio, referred to here as the dehydration volume, is defined to reflect the three-dimensional extent of extremely cold tropopause temperatures. Cold QBO anomalies, coupled with the cold phase of the annual cycle in tropopause temperatures, enhance DV. In controlling the dehydration of tropospheric air to stratospheric water vapor mixing ratios, La Niña events have stronger potential than El Niño events. Dehydration regions with extremely cold temperatures and large sizes occur when cold temperature anomalies associated with the QBO arrive at the TTL in wintertime when the TTL is at the coldest phase of the annual cycle under La Niña conditions.

Acknowledgments. This work was supported by NASA Grant NAS 022490 to the State University of New York at Stony Brook.

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