Diurnal variability of lower and middle atmospheric water vapour over the Asian summer monsoon region: first results from COSMIC-1 and TIMED-SABER measurements

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Abstract
First observations on the vertical structure of diurnal variability of water vapour in the lower and middle atmosphere using 13 years of COSMIC-1 and 18 years of SABER observations over the Asian summer monsoon region are presented in this paper. The most significant and new observation is that the middle stratospheric water vapour (SWV) enhancement is observed between 9 and 18 LT, whereas it is between 6 and 15 LT near tropopause in all the seasons. The diurnal amplitude of water vapour near tropopause is between 0.3 and 0.4 ppmv. Bimodal peaks are found in the diurnal amplitude of SWV, maximizing between 25 and 30 km (~ 0.4 ppmv), and between 45 and 50 km (~ 0.6 ppmv). The analysis reveals that the diurnal variability in the lower SWV is controlled by the tropical tropopause temperature, whereas the middle and upper SWV is primarily controlled by methane oxidation. The results are presented and discussed in the light of present understanding.

Keywords Diurnal variability · Water vapour · COSMIC-1 · TIMED-SABER

1 Introduction
Water vapour (H2O) (WV) is one of the most important greenhouse gases which play a vital role throughout the Earth’s atmosphere in both chemistry and global radiative balance, especially in the upper troposphere and lower stratosphere (UTLS). Dessler et al. (2013) have estimated the climate feedback of about +0.3 Wm−2 K−1 due to an increase in the stratospheric water vapour (SWV), which is responsible for about 5–10% of global warming from all the greenhouse gases. The primary source of hydroxyl ion (OH) in the Earth’s atmosphere is WV, and these OH-radicals control the lifetime of a shorter-lived greenhouse gas like ozone (O3) and long-lived methane (CH4) (Seinfeld and Pandis 2006 and references therein). Any variations from hourly to annual scale in the WV distribution have a significant impact on the global climate-weather system. For example, Forster and Shine (1999) observed that any increase in SWV acts to cool the stratosphere but warm the troposphere. Thus, the SWV plays an important role in stratospheric ozone chemistry and therefore, the global radiation budget.

In general, SWV increases with height due to the slow oxidation of CH4 at 25 km or higher, where stratospheric air is relatively aged (Waugh and Hall 2002). Mote et al. (1996) established the fact that the large seasonal variation of SWV is associated with the tropical cold-point tropopause (CPT) temperature (CPT-T) which is named as “tape recorder effect”. It is agreed that SWV is primarily controlled by the freeze-drying of air passing through the tropical CPT under the influence of the mean upward Brewer-Dobson circulation (BDC) (Dessler et al. 2013). In addition to the CPT temperature, other potential factors control the entry of WV to the lower stratosphere. Overshooting convection can significantly hydrate the lower stratosphere (Uma et al. 2014; Dessler et al. 2016) but overall it is less than 2% (Schoeberl et al. 2018). Besides, small-scale waves and the microphysical processes can also hydrate UTLS (Fueglistaler et al. 2013). In addition to these processes, Quasi-Biennial Oscillation (QBO) also controls the entry of WV to the lower
stratosphere (Dessler et al. 2014; Das and Suneeth, 2020). Once the WV enters the lower stratosphere, it further propagates both vertically and quasi-horizontally (Rosenlof et al. 1997). The CPT temperature is influenced by convection, El Niño-Southern Oscillation (ENSO), and QBO (e.g., Suneeth et al., 2017; Suneeth and Das, 2020). Further, the CPT temperature is controlling the seasonal, annual, and interannual variabilities of SWV (Dessler et al. 2014; Suneeth et al. 2017 and reference therein). Jaing et al. (2015a) have made a detailed study of seasonal and interannual variations of SWV using Aura Microwave limb sounder (MLS) observations. Solomon et al. (2010) discussed the contribution of SWV to the decadal changes in global warming. Schwartz et al. (2008) characterized the role of Madden–Julian Oscillation in hydrating the UTLS region. Suneeth and Das (2020) discussed the influence of Walker circulation in the zonal asymmetry of WV.

Diurnal variation, in general, is associated with the large and well-defined cycle in solar heating during the day and is accounted as the most fundamental component for the variability of the climate system (Gettelman and Rood 2016). The strong coupling between WV and temperature, even on a diurnal scale can provide the basis for a strong positive WV feedback that amplifies the initial temperature changes induced by other greenhouse gases (e.g., CO₂). The periodic heating of the atmosphere by absorption of solar radiation (diurnal-annual) by WV will change the thermal structure, which further can alter the circulation and regulate climate (Sherwood et al. 2010). Diurnal variation of tropospheric WV is also related to moist convection and precipitation, surface wind convergence, and surface evaporation (Dai and Wang 2002). It is to be noted that the diurnal variability of upper tropospheric water vapour is also linked to the diurnal variability of ice clouds (Jian et al. 2015b). Tian et al. (2004) studied the diurnal variability of upper tropospheric WV using geostationary satellite and Chung et al. (2013) with reanalysis. Diurnal variability of relative humidity in the troposphere is reported by Uma and Das (2016), and Emmanuel et al. (2018) using microwave sounder and radiosonde, respectively. Earlier studies have found that the diurnal cycle of tropical lower tropospheric WV lagged the variation in cloud cover by 2 h. It is also observed that WV and upper tropospheric clouds are in-phase over land, whereas it is 12 h out-of-phase over the Oceanic regions (Soden 2000). This is attributed to the vertical structure of clouds over the land and ocean. Even though there are several studies on the diurnal variability of tropospheric WV (Soden 2000; Tian et al. 2004; Uma and Das 2016), as far as the author’s knowledge there is no study reported so far on the diurnal variability of SWV, which is very crucial for understanding climate change and forcing by absorbing solar short-wave radiation.

The maximum amount of WV and other pollutants are injected into the stratosphere over the Asian summer monsoon (ASM) region (Fu et al. 2006). Earlier studies have shown that the maximum amount of WV is accumulated in the UTLS over the ASM anticyclone (Tibetan Plateau) during boreal summer (Randel and Park, 2006; Das and Suneeth 2020). Nützel et al. (2019) found the ASM region that accounts for 14% of SWV in the tropics using a chemistry-transport model. It is also estimated that the ASM region can contribute 75% of the total net upward WV flux in the tropics from July to September (Gettelman and Kinnison 2004). The role of summertime convection over the ASM region on SWV was studied by Wright et al. (2011). Santee et al. (2017) quantified the impact of ASM anticyclone on the climatological seasonal evolution in tracers of UTLS regions. Thus, the region of interest for the present study is the ASM region (Equator to 40°N and 60–100°E), which is assumed to have the alternation of dry and wet seasons due to the seasonal reversal of the monsoon circulation features (Webster et al. 1998). The main intent of this paper is to present and discuss the diurnal variability of WV in the lower and middle atmosphere in the ASM region, which is the first of its kind and its linkage to the diurnal variability of CPT temperature and methane oxidation using satellite measurements.

2 Methodology and data analysis

We used WV and temperature profiles measured by the Constellation Observing System for Meteorology, Ionosphere, and Climate (COSMIC-1) mission based on Global positioning System-Radio Occultation measurement and Sounding of the Atmosphere using Broadband Emission Radiometry (SABER) onboard Thermosphere Ionosphere Mesosphere Energetics Dynamics (TIMED) satellite. Temperature and WV from COSMIC-1 are obtained from moist air retrieval (wetPrf, level-2) (Kursinski et al. 1997) during 2006–2019. In SABER, temperature and WV are derived from 15 and ~ 6.8 µm, respectively during 2001–2019 (Remsberg 2008; Rong et al. 2019). The vertical resolution of COSMIC-1 is 100 m and SABER is 1 km. A thorough quality check is done for both temperature and water vapour by removing the wild points and outliers. Since the temperature follows the normal distribution, we estimated monthly mean and ± 3σ (where σ is the standard deviation) and then removed the temperature data, which falls outside monthly mean ± 3σ (outliers). A typical example for quality control of temperature for January 2007–2019 is shown in Fig. S1. Water vapour distributions are skewed towards the higher values, thus we have estimated 0.25 and 99.75 percentiles, and the data points which fall outside these percentile values are removed in the analysis. A typical example for quality control of water vapour is shown in Fig. S2. In this process, we obtained > 90% of useful data. Diurnal composite
profiles (every 3 h) for each month with a grid resolution of 5° (latitude) × 10° (longitude) were generated and each grid has 30–40 and 60–75 profiles/month for COSMIC-1 and SABER, respectively. We have carried out the significant test (student’s t-test) and found that the analysis of unevenly distributed data does not significantly affect our results due to the large sample size.

The COSMIC-1 derived temperature and WV are extensively validated with radiosonde, Aura-MLS, Atmospheric Infrared Sounder (AIRS), and different reanalysis data (Sun et al. 2010; Kishore et al. 2011; Rieckh et al. 2018). Individual WV profile from COSMIC-1 is accurate up to 8 km and seasonal mean up to 10 km with a fractional difference of 5–10%. The COSMIC-1 temperature has an accuracy of ~0.5 K for individual profiles (Kishore et al. 2009). Alexander et al. (2014) have taken 20 days of data with a grid of 20° × 40° (latitude × longitude) over two regions (30–50°N, 160–200°W and 30–50°N, 80–120°W) and found an average bias of 0.1 K. Rong et al. (2019) validated the WV from SABER and estimated a total systematic error of 10–20%. Further, the authors have compared SABER measured WV with other satellite measurements, which shows a good agreement in the seasonal mean profile within ±10%. SABER temperature is well compared with other measurements and has a precision of ~1 K (Remsberg et al. 2003). Yue et al. (2019) have utilized the data of WV from SABER for trend analysis.

### 3 Results and discussions

#### 3.1 Climatology of WV

Figure 1 shows the seasonal mean distribution of WV mixing ratio (WMR) at 6, 8, 16, 25, 40, and 50 km for four seasons, i.e. December–January–February (DIF), March–April–May (MAM), June–July–August (JJA), and September–October–November (SON). The closed contour during JJA at 16 km indicates the climatological location of ASM-anticyclone following Park et al. (2007). The maximum amount of upper tropospheric WV is found over the southeast of ASM-anticyclone during JJA. WV at 6 and 8 km are high over the Tibetan Plateau during JJA. At 16 km, the WV is high (6–8 ppmv) between the equator and 15°N; however, during JJA and SON the increase extended up to 40°N. Monsoon season acts as one of the significant sources of WV, and several studies discussed the transport of WV into the UTLS region (Randel et al. 2015). Above 16 km, WV is found to be very less, and between 25 and 40 km, it is about 4–5 ppmv and does not have drastic seasonal variations as observed in the troposphere. At 40 km, WV is high (~7 ppmv) in mid-latitude regions compared to the tropics, especially during DIF.

At 50 km over the entire ASM region, the WV is about 7 ppmv during all seasons. The seasonal characteristics of WV measured by COSMIC-1 and SABER are similar to the other satellite measurements as reported earlier. For example, both SABER and MLS show a similar distribution of WV (6–8 ppmv) near the tropopause (~16 km or 100 hPa) (Randel and Park 2006).

#### 3.2 Diurnal variability of WV and temperature

We divide the ASM into three regions: Region 1 (R1), Region 2 (R2), and Region 3 (R3) based on the intensity of convection. R1 covers the deep-equator from Equator–10°N, R2 is the central part of ASM, which is the monsoon active region from 15 to 25°N, and R3 covers ASM-anticyclone from 30 to 40°N. Figure 2 shows the diurnal variability of WV for different seasons/regions from 2 to 65 km. The mean removed diurnal variability is shown in Fig. 3 and its standard errors in Figure S3 (supplementary figure). We also plotted seasonal time-series of diurnal variability of WV at 6, 8, 16, 25, 40, and 50 km as shown in Figure S4 to see the inter-annual consistency. The maximum extent of WV is observed during JJA for all regions. Strong diurnal variability is observed (Fig. 3) in all the regions/seasons and has inter-annual consistency (Fig. S4). All the regions/seasons show the variability of WV within ±400 ppmv (~1 ppmv) at 2–10 km (15–65 km) and have phase propagation similar to tidal waves. In the troposphere (<16 km), maximum WV is observed during 9–18 LT. WV in the UTLS region is also found to be high (~0.8 ppmv) during 9–18 LT. Similarly, enhanced WV is observed during 9–18 LT between 25 and 35 km. However, it is also to be noted that the local time at which the diurnal peak occurs is not always identical between regions and seasons, which is attributed to the availability of water vapour and the cold point temperature. Slow CH₄ oxidation takes place between 25 and 35 km by the following reactions (Seinfeld and Pandis 2006):

\[
\text{CH}_4 + \text{OH} \rightarrow \text{CH}_3 + \text{H}_2\text{O} \quad \text{(R1)}
\]

\[
\text{CH}_4 + \text{O}^{(1)}\text{D} \rightarrow \text{CH}_3 + \text{OH} \quad \text{(R2)}
\]

CH₄ is also long-lived species in the stratosphere, which is transported from the surface and found to be ~1.2 ppmv in the stratosphere. When the intensity of ultraviolet (UV) radiation increases from forenoon to afternoon, the WV in the stratosphere reacts with ozone between 25 and 35 km (maximum ozone) to form OH-radicals by the following reactions R3, R4, R5, R6. The key production of OH in the stratosphere is initiated by photolysis of O₃ to produce O(1D) by the following reaction (Seinfeld and Pandis 2006):
The total rate of OH production between 20 and 50 km from the reactions \( R4 \) and \( R5 \) is \( \sim 1.8 \times 10^4 \) molecules cm\(^{-3}\) s\(^{-1}\) and \( 0.2 \times 10^4 \) molecules cm\(^{-3}\) s\(^{-1}\) (Seinfeld and Pandis 2006).

These OH-radicals form during the daytime and maximize to higher heights are derived from TIMED-SABER from January 2002 to December 2019. Black closed contour at 16 km during JJA indicates the climatological location of the Asian summer monsoon (ASM) anticyclone, which is defined by Park et al. (2007) as a stream function contour (400 m\(^2\) s\(^{-1}\)) (for more details please refer Park et al. 2007 and Santee et al. 2017).

\[
\begin{align*}
O_3 + h\nu & \rightarrow O_2 + O(^1D) & (R3) \\
O(^1D) + H_2O & \rightarrow 2OH & (R4) \\
O(^1D) + CH_4 & \rightarrow OH + CH_3 & (R5) \\
H_2O + h\nu & \rightarrow OH + H & (R6)
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Thus, the oxidation of one CH$_4$ molecule yields exactly two molecules of H$_2$O in the lower and middle stratosphere. The rate coefficient ($k_1$) for reaction (1) and is given by equation (E1) (Seinfeld and Pandis 2006)

$$k_1 = 2.45 \times 10^{-12} e^{-\frac{1775}{T}} \text{ cm}^3 \text{ molecule}^{-1} \text{ s}^{-1}$$

(E1)

where T is the temperature and thus at 230 K, $k_1_{230K} = 1.09 \times 10^{-15} \text{ cm}^3 \text{ molecule}^{-1} \text{ s}^{-1}$.

CH$_4$ mixing ratio VMR$_{CH4}$ is about 1.2 ppmv in the stratosphere (~30 km) (e.g., Myhre et al. 2007) and thus,

Fig. 2 Diurnal variability of WMR for different seasons of (a) Region 1 (R1), (b) Region 2 (R2), and (c) Region 3 (R3) (see text for region averaging). COSMIC-derived water vapour is from 2 to 10 km, whereas SABER derived is from 15 to 65 km.
we obtained a concentration \(n\) of \(\text{CH}_4\) about \(4.5 \times 10^{11}\) molecules \(\text{cm}^{-3}\) using a simple equation (E2).

\[
n = 1.09 \times 10^{-12} \times \text{VMR}_{\text{CH}_4} \times p/(k \times T)\]  

(E2)

Here, the stratospheric pressure \((p)\) is taken as 1200 Pa (at 30 km) and the Boltzmann constant \(k = 1.3807 \times 10^{-23}\) J-K \(\text{mol}^{-1}\). OH concentration is found to be \(2.5 \times 10^6\) molecules \(\text{cm}^{-3}\) (e.g., Seinfeld and Pandis, 2006 and Min-schwaner et al., 2011). Further, we estimate the production rate of \(\text{H}_2\text{O}\) due to methane oxidation (Reaction 1) using the equation (E3) and found to be about \(1.2 \times 10^3\) molecules \(\text{cm}^{-3} \text{s}^{-1}\).
A modelling study by Zahn et al. (2006) also shows net \( H_2O \) production due to methane oxidation is \( \sim 2500 \) molecules cm\(^{-3}\) s\(^{-1}\). The estimated water vapour from methane oxidation is \( \sim 0.4 \) ppmv per 24 h between 25 and 35 km, which is consistent with the observations. Thus the observed enhancement of WV between 25 and 35 km during 9–18 LT is due to the availability of OH-radicals. The enhanced water vapour above 35 km is observed much earlier (3–6 LT) than the lower height due to the availability of UV radiation over that height range. During early morning hours, the solar zenith angle lies between 90 and 110 degree, and the observer will have no UV radiation in the troposphere but it will reach a higher height due to the Sun’s position. At 45 km, WV is maximizing (~0.8 ppmv) at 0–6 LT, and at 60 km it is during 6–9 LT.

Diurnal mean removed temperature for different regions/seasons is shown in Fig. 4 and its standard error in Fig. S5 (Supplementary Fig.). Diurnal variability is \( < \pm 0.5 \) K in the troposphere, whereas in the stratosphere it is about \( \pm 2 \) K for all the regions. The lower troposphere and between 35 and 60 km are warmer for all regions during 12–18 LT and more during JJA. The UTLS region is warmer during 6–12 LT for all regions/seasons. This pattern weakens during SON. Distinct semi-diurnal variation is observed above 35 km and up to 65 km, showing the presence of semi-diurnal tides for all the regions/seasons. There are many studies on diurnal tides using temperature profiles measured from COSMIC-1 (e.g., Xie et al. 2010) and SABER (e.g., Sakazaki et al. 2018), thus we are not discussing further.

**3.3 Correlation between water vapour and temperature**

To get further insight into the variability and the relationship between WV and temperature, we have shown the time-series (absolute) plots at 6, 8, 16, 25, 40, and 50 km for different regions (R1, R2, and R3) and seasons as shown in Fig. 5a–f. WV variability is in phase with temperature between 6 and 16 km for all regions/seasons. The holding capacity of WV in dry air depends on the saturation mixing ratio (SMR) and thus WV and temperature are in phase. However, we observed a phase lag of 3 h between temperature and WV for R3, which could be due to deep-convection over this region. Deep-convection has warm anomalies (<1 K) between 10 and 14 km as well as above CPT (Johnston et al. 2018). Since WV lag deep-convection by 2 h over land (Soden 2000), there is a phase lag between temperature and WV over R3. The deep convection generates cirrus anvil clouds that can further moisten the upper troposphere by the evaporation of the cirrus anvil. Thus, the temperature near the tropopause, overshooting convection, and cirrus anvil are the few factors that control the diurnal variation of WV in the UTLS region.

WV and temperature are always out-of-phase above 25 km in all the regions/seasons (Fig. 5d). As discussed earlier, methane oxidation occurs between 25 and 35 km. The maximum WV occurs during 9–18 LT in all regions/seasons, which is attributed to the availability of the maximum amount of OH-radicals during daytime as discussed above. The phase delay of ~12 h is observed between WV and temperature for R1 and R2, whereas for R3 it is ~6 h. The regional differences can be attributed to the spatial change in the thermal differences. Above 30 km (Fig. 5e, f) WV and temperature show opposite phases for all seasons/regions and the maximum WV is observed during 0–9 LT, maximizing at 6 LT. This may be due to the availability of OH radical in the presence of UV radiation, which is available much earlier in the morning. The maximum temperature is observed during 12–18 LT. Both WV and temperature show distinct diurnal variation indicating the influence of diurnal tides with a phase difference of 12 h.

Figure 6 shows the CPT-T and altitude (CPT-A) for different seasons for R1 and R2 along with the diurnal mean removed WV and saturation mixing ratio (SMR). We used lapse rate tropopause temperature (LRT-T) and altitude (LRT-A) for R3, as we cannot detect CPT except for JJA. The tropopause is warmer by 0.6–1 K during 6–15 LT for all seasons for R1 and R2 and we also observed maximum WV during 9–18 LT. The tropopause altitude varies between 16.8 to 17.5 km for R1 and R2. Similar warmer tropopause and high WV are found for R3, but during DJF and MAM. During JJA and SON in R3, the tropopause is warmer during 0–9 LT and high WV is observed during 9–15 LT. LRT-A varies between 11 and 15 km for R3, maximizing during JJA, which is in contrast to R1 and R2 (Das and Suneeth 2020). SWV is strongly coupled to the tropopause temperature (e.g., Randel et al. 2013) and thus the annual cycle of the tape-recorder effect is observed (Mote et al. 1996). In addition to the annual cycle, the present analysis show that the diurnal variability of tropopause temperature is also controlling the lower SWV.

Further, we correlate the temperature and WV for all regions/seasons as shown in Fig. 7. A high correlation is observed between temperature and WV near the tropopause, indicating that tropopause temperature controls the SWV, which is more or less consistent for all regions and seasons. We also observed a high correlation between WMR and temperature over the altitude of 30–35 km, where methane oxidation takes place as discussed above. Thus, the temperature in the UTLS region, and between 25 and 35 km is the controlling factor for SWV.
3.4 Diurnal amplitude and phase

Further, the diurnal amplitude and phases are extracted for both WV and temperature for all regions/seasons as shown in Fig. 8. For WV (temperature), diurnal amplitude and phase are extracted between 5 and 10 km (5 and 35 km) from COSMIC-1 and 15–60 km (15–60 km) from SABER. The amplitude of WV is high in the lower troposphere which is well documented in the literature (e.g., Soden 2000; Tian et al. 2004; Uma and Das 2016), thus we will emphasize more discussion on diurnal variability of WV near the tropopause and the entire stratosphere. WV amplitude of about 0.3–0.4 ppmv is found near the tropopause. Over R1, we observed bimodal peaks, one at 25–30 km (0.4 ppmv) and the other at 45–55 km (0.6 ppmv). The bimodal peak is also observed in R2 and R3 in line with R1, but the second peak occurs between 40 and 45 km. It is interesting to note that the amplitude of both the peaks in R2 is the same and found
to be 0.4 ppmv. Over R3, the amplitude of the primary peak (25–30 km) is more than the secondary peak. The characteristic of amplitude in WV is almost similar in all the seasons except the occurrence of the secondary peak, which varies within 5 km. Forster and Shine (1999) have shown a strong radiative forcing due to the increase of SWV. They estimate that a 0.7 ppmv increase in SWV would result in the radiation forcing of 0.19–0.29 Wm$^{-2}$. We estimated a net radiative forcing of ~0.4 Wm$^{-2}$ for the diurnal variability in SWV of 0.4–0.6 ppmv using SBDART (Ricchiazzi et al. 1998). Thus the diurnal component of SWV also influences the Earth’s radiation budget. Random phase profiles in WV are observed below the tropopause and above it to 35 km, where we observed a phase of 12–16 h. Downward phase propagation

Fig. 5 Diurnal variability of WMR (blue) and temperature (T) (red) at (a) 6 and (b) 8 km from COSMIC, and (c) 16, (d) 25, (e) 40 and (f) 50 km from SABER measurements for different seasons for R1, R2, and R3. Vertical lines indicate the respective standard errors.
is observed from 4 to 8 h between 35 and 45 km. Like amplitude, all seasons show similar phase profiles in WV. Diurnal amplitude in temperature shows four peaks, viz. 16–18 km (near tropopause), 30–35 km, 45–50 km and 60 km. Clear downward phase propagation in temperature is observed between 10 and 40 km in all regions/seasons. There is no phase propagation above 40 km. Diurnal amplitude and phase profiles in temperature are well established (Xie et al. 2010; Sakazaki et al. 2018). The common amplitude and phase profiles of temperature (15–35 km) derived from COSMIC-1 and SABER show the same characteristics. Both the peaks in WV and temperature in R1 coincide, however, the secondary peak slightly varies within 5 km. Over R2 and R3, there is no significant amplitude peak in temperature below 35 km. However, the enhanced amplitude is observed

Fig. 5 (continued)
between 45 and 55 km over R2 and R3 for all the seasons. Even though the diurnal amplitude of temperature is very less near the tropopause, we observe enhanced diurnal amplitude in WV. Between 25 and 35 km and between 45 and 55 km, both temperature and WV show enhanced diurnal amplitude with ± 5 km variability in the peak altitude.

### 4 Concluding remarks

In this study, the vertical structure of the diurnal variability of SWV and its controlling factors are presented. We find the diurnal variability of lower SWV is mainly controlled by the tropopause temperature, whereas in the middle and upper stratosphere it is mainly controlled by

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**Fig. 5 (continued)**
methane oxidation in addition to temperature. We also show the amplitude and phase profiles of diurnal variation of water vapour and temperature for different seasons. Bimodal and trimodal distributions of diurnal amplitude in water vapour are observed. A systematic phase difference is noted between the water vapour and temperature, which is very consistent in the stratosphere. The present study has shown the diurnal variability of SWV over the ASM region, which is the first of its kind. Further investigations are essential to evaluate the role of small-scale waves and turbulence occurring in the upper troposphere and lower stratosphere region to have a complete understanding of the observations presented here.

![Fig. 6](image)

Diurnal variability of mean removed WMR (black) and saturation mixing ratio (SMR) (red) at 17–18 km derived from SABER, tropopause altitude (blue), tropopause temperature (magenta) derived from COSMIC temperature profiles for different seasons of (a) R1, (b) R2, and (c) R3. Cold-point tropopause altitude (CPT-A) and temperature (CPT-T) are estimated for R1 and R2, whereas lapse rate tropopause (LRT) is estimated for R3, as well-defined CPT cannot be obtained for R3, except for JJA. Vertical lines indicate the respective standard errors.
Fig. 7  Height profiles of correlation coefficient (r) between WMR and T for different seasons of (a) R1, (b) R2, and (c) R3. The profiles from 5 to 10 km (5–25 km) are from COSMIC derived WMR (T), whereas the profile from 15 to 60 km is from SABER derived WMR and T.

Fig. 8  Height profiles of diurnal amplitude and phase of (a) WMR, and (b) T for different seasons at R1, R2, and R3. The profiles from 5 to 10 km (5–25 km) are from COSMIC derived WMR (T), whereas the profile from 15 to 60 km is from SABER derived WMR and T.
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Data availability  The data used in the present study are downloaded from the following websites: SABER: http://saber.gats-inc.com/data.php. COSMIC-1: http://cdiac-www.cosmic.ucar.edu/cdiaac/index.html.

Declarations

Conflict of interest  The authors declare that there is no conflict of interest.

References

Alexander P, de la Torre A, Llamado P, Hierro R (2014) Precision estimation in temperature and refractivity profiles retrieved by GPS radio occultations. J Geophys Res : Atmos 119(14):8624–8638. https://doi.org/10.1002/2013JD021016

Chung ES, Soden BJ, Sohn BJ, Schmetz J (2013) An assessment of the diurnal variation of upper tropospheric humidity in reanalysis data sets. J Geophys Res 118(9):3425–3430. https://doi.org/10.1002/jgrd.50345

Dai A, Wang J, Ware RH, Van Hove T (2002) Diurnal variation in water vapor over North America and its implications for sampling errors in radiosonde humidity. J Geophys Res. https://doi.org/10.1029/2001JD000642

Das SS, Suneeth KV (2020) Seasonal and interannual variations of water vapor in the upper troposphere and lower stratosphere over the Asian Summer Monsoon region-in perspective of the tropopause and ocean-atmosphere interactions. J Atmos Sol-Terr Phys. https://doi.org/10.1016/j.jastp.2020.105244

Dessler AE, Schoeberl MR, Wang T, Davis SM, Rosenlof KH (2013) Stratospheric water vapor feedback. Proc Natl Acad Sci USA 110(45):18087–18091. https://doi.org/10.1073/pnas.1310344110

Dessler AE, Schoeberl MR, Wang T, Davis SM, Rosenlof KH, Vernier JP (2014) Variations of stratospheric water vapor over the past three decades. J Geophys Res 119(22):12–588. https://doi.org/10.1002/2014JD021712

Dessler AE, Ye H, Wang T, Schoeberl MR, Oman LD, Douglass AR et al (2016) Transport of ice into the stratosphere and the humidification of the stratosphere over the 21st century. Geophys Res Lett 43(5):2323–2329. https://doi.org/10.1002/2016GL067991

Emmanuel M, Sunilkumar SV, Ratnam MV, Muhsin M, Parameswaran K, Murthy BK (2018) Diurnal variation of the tropospheric water vapour over a coastal and an inland station in Southern Indian Peninsula. J Atmos Sol-Terr Phys 179:11–21. https://doi.org/10.1016/j.jastp.2018.06.007

Forster PMDF, Shine KP (1999) Stratospheric water vapour changes as a possible contributor to observed stratospheric cooling. Geophys Res Lett 26(21):3309–3312. https://doi.org/10.1029/1999GL010487

Fu R, Hu Y, Wright JS, Jiang JH, Dickinson RE, Chen M, Filipiak M, Read WG, Waters JW, Wu DL (2006) Short circuit of water vapor and polluted air to the global stratosphere by convective transport over the Tibetan Plateau. Proc Natl Acad Sci USA 103(15):5664–5669. https://doi.org/10.1073/pnas.0601584103

Fueglistaler S, Liu YS, Flannaghan TJ, Haynes PH, Dee DP, Read WJ, Bernath PF (2013) The relation between atmospheric humidity and temperature trends for stratospheric water. J Geophys Res : Atmos 118(2):1052–1074. https://doi.org/10.1002/2012JD019517

Gettelman A, Rood RB (2016) Components of the Climate System. In: Demystifying climate models. Earth systems data and models, vol 2. Springer, Berlin, Heidelberg. Series editors: Bernd Blasius, Carl von Ossietzky University Oldenburg, Oldenburg, Germany; William Lahoz, NILU—Norwegian Institute for Air Research. Kjeller, Norway; Dimitri P. Solomatine, UNESCO—IHE Institute for Water Education, Delft, The Netherlands. https://doi.org/10.1007/978-3-662-48959-8_2

Gettelman A, Kinnison DE, Dunkerton TJ, Brasseur GP (2004) Impact of monsoon circulations on the upper troposphere and lower stratosphere. J Geophys Res. https://doi.org/10.1029/2004JD004878

Jiang JH, Su H, Zhai C, Wu L, Minschwaner K, Molod AM, Tompkins AM (2015a) An assessment of upper troposphere and lower stratosphere water vapor in MERRA, MERRA2, and ECMWF reanalyses using Aura MLS observations. J Geophys Res 120(11):11468–11485. https://doi.org/10.1002/2015JD023752

Jiang JH, Su H, Zhai C, Shen TJ, Wu T, Zhang J et al (2015b) Evaluating the diurnal cycle of upper-tropospheric ice clouds in climate models using SMILES observations. J Atmos Sci 72(3):1022–1044. https://doi.org/10.1175/JAS-D-14-0124.1

Johnston BR, Xie F, Liu C (2018) The effects of deep convection on regional temperature structure in the tropical upper troposphere and lower stratosphere. J Geophys Res 123(3):1585–1603. https://doi.org/10.1002/2017JD027120

Kishore P, Namboothiri SP, Jiang JH, Sivakumar V, Igarashi K (2009) Global temperature estimates in the troposphere and stratosphere: a validation study of COSMIC/FORMOSAT-3 measurements. Atmos Chem Phys 9:897–908. https://doi.org/10.5194/acp-9-897-2009

Kishore P, Ratnam MV, Namboothiri SP, Velicogna I, Basha G, Jiang JH, Igarashi K, Rao SVB, Sivakumar V (2011) Global (50 degrees S-50 degrees N) distribution of water vapor observed by COSMIC GPS RO: Comparison with GPS radiosonde, NCEP, ERA-Interim, and JRA-25 reanalysis data sets. J Atmos Sol-Terr Phys. https://doi.org/10.1016/j.jastp.2011.04.017

Kurinskius ER, Hajj GA, Schofield JT, Linfield RP, Hardy KR (1997) Observing Earth’s atmosphere with radio occultation measurements using the global positioning system. J Geophys Res 102(D19):23429–23465. https://doi.org/10.1029/97JD01569

Minschwaner K, Manney GL, Wang SH, Harwood RS (2011) Hydroxyl in the stratosphere and mesosphere-part 1: diurnal variability. Atmos Chem Phys 11(3):955. https://doi.org/10.5194/acp-11-955-2011

Mote PW, Rosenlof KH, McIntyre ME, Carr ES, Gille JC, Holton JR, Waters JW (1996) An atmospheric tape recorder: the imprint of tropical tropopause temperatures on stratospheric water vapor. J Geophys Res 101(D2):3989–4006. https://doi.org/10.1029/95JD03422

Myhre G, Nilsen JS, Gulstad L, Shine KP, Rognerud B, Isaksen IS (2007) Radiative forcing due to stratospheric water vapour from CH4 oxidation. G Res Lett 34(1):L01807. https://doi.org/10.1029/2006GL027472

Nützel M, Podglajen A, Garny H, Ploeger F (2019) Quantification of water vapour transport from the Asian monsoon to the stratosphere. Atmos Chem Phys. https://doi.org/10.5194/acp-19-8947-2019
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