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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Research Article

Keywords: Diurnal variability, Water vapour, COSMIC-1, TIMED-SABER

Posted Date: August 17th, 2021

DOI: <https://doi.org/10.21203/rs.3.rs-693267/v1>

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Version of Record: A version of this preprint was published at Climate Dynamics on April 22nd, 2022. See the published version at <https://doi.org/10.1007/s00382-022-06282-5>.

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 tropospheric water vapour in the lower and middle atmosphere using 13 years of COSMIC and 18 years of SABER observations are presented in this paper. The most significant and new observation is that the middle stratospheric water vapour (SWV) enhancement is observed between 9-18 LT, whereas it is between 6-15 LT near tropopause in all the seasons. The diurnal amplitude of water vapour near tropopause is between 0.3-0.4 ppmv. Bimodal peaks are found in the diurnal amplitude of SWV, maximizing between 25-30 km (~0.4 ppmv) and 45-50 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 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 (H_2O) (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 $+0.3 \text{ Wm}^{-2} \text{ K}^{-1}$ due to an increase in the stratospheric water vapour (SWV), which is about 5-10% responsible for the 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 (O_3) and long-lived like methane (CH_4) (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 CH_4 , 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 process can also hydrate UTLS (Fueglistaler et al.

2013). It is observed that stronger convection leads to a relatively dry stratosphere, whereas weaker convection leads to a wetter stratosphere (Randel et al. 2015). 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). Many studies that deal with the seasonal, annual, and interannual variabilities of SWV, which are further linked with the CPT temperature which is in principle controlled by convection, El Niño - Southern Oscillation (ENSO), and QBO (Dessler et al. 2014; Suneeth et al. 2017 and reference therein). 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. In addition to it, the entry of WV into the stratosphere can be influenced by the overshooting convection, which can hydrate the lower stratosphere (Dessler et al. 2016) and small-scale waves and microphysical process (Fueglistaler et al. 2013).

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 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, which amplify both internal and external climate as well as by latent heat release excites planetary-scale waves that propagate in the whole atmosphere, which can alter the circulation and have a climate impact (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). Tian et al. (2004) studied the diurnal variability of upper tropospheric WV using Geostationary satellite and Chung et al. (2013) with reanalysis. Variability of relative humidity in the diurnal scale is reported by Uma and Das (2016) using microwave sounder and by radiosonde by Emmanuel et al. (2018). Earlier studies have found that the diurnal cycle of tropical lower tropospheric WV lagged the variation in cloud cover by 2 hrs. It is also observed that WV and upper tropospheric clouds are in-phase over land, whereas it is 12 hrs 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), but, as far as the author's knowledge there is no study reported so far in the diurnal variability of SWV, which is very crucial for understanding climate change and forcing by absorbing solar short-wave infrared radiation.

The maximum amount of WV and other pollutants are injected into the stratosphere over the 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 that ASM contributes 14 % of SWV in tropics using a chemistry-transport model. It is also estimated that the ASM can contribute 75 % of the total net upward WV flux in the tropics from July-September (Gettelman and Kinnison, 2004). The role of summertime convection over the ASM on SWV was studied by Wright et al. (2011). Santee et al. (2017) quantified the impact of ASM anticyclone in the climatological seasonal evolution in tracers of UTLS regions. Thus, the region of interest for the present study is ASM (Equator to 40°N and 60-100°E), which assumed to have

the most distinct variation of the annual cycle and the alternation of dry and wet seasons, which is in concert with 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 (from the troposphere to stratosphere) 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) 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 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 $\sim 6.8 \mu\text{m}$, respectively during 2001-2019 (Remsberg 2008; Rong et al. 2019). The vertical resolution of COSMIC 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, thus we estimated monthly mean and $\pm 3\sigma$ (where σ is the standard deviation) and then remove the temperature data, which falls outside monthly mean $\pm 3\sigma$ (outliers). A typical example for quality control of temperature for January 2007-2019 is shown in Figure 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 falls outside these percentile values are removed in the analysis. A typical example for quality control of water vapour is shown in Figure S2. In this process, we obtained $> 90 \%$ of useful data.

Diurnal composite profiles (every 3-hrs) for each month with a grid resolution of 5° (latitude) $\times 10^\circ$ (longitude) were generated and each grid has 30-40 and 60-75 profiles/month for COSMIC and SABER, respectively. We have carried out the significant test (t-student test) and found that the analysis of unevenly distributed data does not significantly affect our results due to the large sample size.

The COSMIC derived temperature and WV are extensively validated with radiosonde, Micro-limb sounder (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 is accurate up to 8 km and mean up to 10 km with a fractional difference of 5-10 %. The COSMIC temperature has an accuracy of ~ 0.5 and 0.1 K for individual and averaged profiles, respectively (Kishore et al. 2009; Alexander et al. 2014). Rong et al. (2019) validated the WV from SABER and Yue et al. (2019) utilized the data set for trend analysis. A total systematic error, which is calculated as the root-sum-square of individual systematic sources is about 10-20% and overall excellent agreement with a mean profile within $\pm 10\%$. SABER temperature is well compared with other measurements and found a precision of ~ 1 K (Remsberg et al. 2003).

3. Results and discussions

3.1. Climatology of WV

Figure 1 shows the seasonal mean distribution of WV mixing ratio (WVR) at 6, 8, 16, 25, 40, and 50 km for four seasons, i.e. December-January-February (DJF), March-April-May (MAM), June-July-August (JJA), and September-October-November (SON). The close contour during JJA at 16 km indicates the climatological locations of ASM-anticyclone following Park et al. (2007). Tropospheric WV is high and maximum is found over the south-east 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 25-40 km, it is about 4-5 ppmv and does not have drastic seasonal characteristics as we observed in the troposphere. At 40 km, WV is high (~7 ppmv) in mid-latitude regions compared to the tropics, especially during DJF. At 50 km over the entire ASM region, the WV is about 7 ppmv during all the seasons. The seasonal characteristics of WV measured by the COSMIC-1 and SABER are similar to the other satellite measurements as reported earlier. For example, near the tropopause (~ 16 km or 100 hPa) both SABER and MLS show WV of 6-8 ppmv with similar distributions (Randel and Park 2006).

3.2. Diurnal variability of WV and temperature

We divide the ASM into three regions: R1, R2, and 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 15-25°N and R3 covers ASM-anticyclone from 30-40°N. Figure 2 shows the diurnal variability of WV for different seasons/regions from 2-65 km. The mean removed diurnal variability shown in Figure 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 (Figure 3) in all the regions/seasons and has inter-annual consistency (Figure 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 25-35 km. Between 25-35 km, the air is relatively aged (Waugh and Hall, 2002), which is attributed to slow CH₄ oxidation by the following reactions (Seinfeld and Pandis 2006):

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 25-35 km (maximum ozone) to form OH-radicals by the following reactions:

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 20-50 km from equation-4 and equation-5 is ~ 1.8x10⁴ molecules cm⁻³s⁻¹ and 0.2x10⁴ molecules cm⁻³s⁻¹ (Seinfeld and Pandis 2006). These OH-radicals form during the daytime and maximize (2x10⁴ molecules cm⁻³) during 10-14 LT and reacts with methane (equation-1) to form WV (Seinfeld and Pandis 2006). The associated rate coefficient for equation-6 is estimated as $k_1 = 1.09 \times 10^{-15} \text{ cm}^3 \text{ molecule s}^{-1}$ at 230 K (~30 km) and thus the production rate of H₂O from methane oxidation (equation-6) is ~1.2 × 10³ molecules cm⁻³ s⁻¹. A modelling study by Zahn et al. (2006) also shows net H₂O production due to methane oxidation is ~2500 molecules cm⁻³s⁻¹. The estimated water vapour from methane oxidation is

~0.4 ppmv per 24 hrs between 25-35 km, which consistent with the observations. Thus the observed enhancement of WV between 25-35 km during 9-18 LT is due to the availability of OH-radicals. At higher altitudes (>35 km), the time of enhancement in the WV decreases with altitude from 3-6 LT, which is attributed to the availability of UV radiation. At 45 km, it 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 Figures 4 and its standard error in Figure S5 (supplementary figure). Diurnal variability is $<\pm 0.5\text{K}$ in the troposphere, whereas in the stratosphere it is about $\pm 2\text{K}$ for all the regions. Lower troposphere and between 35-60 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 Figures 5 a-f. WV variability is in phase with temperature between 6-16 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-hrs between temperature and WV for R3, which could be due to deep-convection over this region. Deep-convection has warm anomalies between 10-14 km as well above CPT ($<1\text{ K}$) (Johnston et al. 2018). Since WV lag deep-convection by 2 hr

overland (Soden 2000), thus 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 (Figure S8 d). As discussed earlier, methane oxidation occurs between 25-35 km. The maximum WV occurs during 9-18 LT in all regions/seasons, which is attributed to the availability of OH-radicals during day-time as discussed above. The phase delay of ~ 12 is observed between WV and temperature for R1 and R2, whereas for R3 it is ~ 6 hr. Above 30 km (Figures 5 e-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 early 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 hrs.

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 DJF and MAM. During JJA and SON in R3, the tropopause is warmer during 0-9 and high WV is observed during 9-15 LT. LRT-A varies between 11-15 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), which is not only restricted to the annual cycle (like tape-recorder effect, Mote et al. 1996) but also to the diurnal scale as observed in the present analysis. Thus more amount of WV in the lower stratosphere is due to the tropopause temperature.

Further, we correlate the temperature and WV for all regions/seasons as shown in Figure 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 30-35 km, where methane oxidation takes place as discussed above. Thus, the temperature in UTLS region and between 25-35 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 Figure 8. For WV (temperature), diurnal amplitude and phase are extracted between 5-10 km (5-35 km) from COSMIC 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 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 height 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 0.7 ppmv increase in SWV would result in a radiation forcing of 0.19-0.29 Wm^{-2} . We estimate the net radiative forcing of $\sim 0.4 \text{ Wm}^{-2}$ for the diurnal variability SWV of 0.4-0.6 ppmv using SBDART (Ricchiazzi et al. 1998). Thus the diurnal component of SWV has a good impact on the Earth's radiation budget.

The phase is random in the troposphere, whereas from the tropopause to 35 km, the phase is during 12-16 hrs in all three regions. Downward phase propagation is observed from 4 to 8 hrs between 35-45 km. Like amplitude, all seasons show similar phase profiles. Diurnal amplitude in temperature shows four peaks, viz. 16-18 km (near tropopause), 30-35 km, 45-50 km and 60 km. 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 between 45-55 km over R2 and R3 for all the seasons. Clear downward phase propagation 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 and SABER shows the same characteristics. Even though the amplitude of diurnal amplitude of temperature is very less near the tropopause, but 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. Both temperature and WV show similar downward phase propagation.

4 Concluding remarks

In this study, the vertical structure of the diurnal variability of SWV and its controlling factor is 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 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 a trimodal distribution 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 an understanding of the observations presented here.

Data availability

The data used in the present study are downloaded from the following websites:

SABER: <http://saber.gats-inc.com/data.php>

COSMIC: <http://cdaac-www.cosmic.ucar.edu/cdaac/index.html>.

Declaration of competing for interest

The authors declare that there is no conflict of interest.

Acknowledgments

The authors are greatly appreciative to the teams of SABER-TIMED and the COSMIC Data Analysis and Archive Centre, and sincerely acknowledged them for providing the data used in this study.

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Figure Captions

Figure 1. Seasonal mean latitude-longitude distribution of water vapour at 6, 8, 16, 25, 40, and 50 km for December-January-February (DJF), March-April-May (MAM), June-July-August (JJA), and September-October-November (SON). Water vapour mixing ratio up to 8 km is measured using COSMIC-1 from December 2006 to November 2019, and from 16 km onwards to higher levels are derived from TIMED-SABER from January 2002 to December 2019. Black closed contour (at 6, 8 and 16.5 km) indicates the climatological location of the Asian summer monsoon (ASM) anticyclone during JJA as defined by Park et al., (2007) by defining a stream function contour ($400 \text{ m}^2 \text{ s}^{-1}$) (for more detail please refer Park et al. 2007 and Santee et al. 2017).

Figure 2. Diurnal variability of water vapour mixing ratio (WMR) for different seasons of Region-1 (R1), Region-2 (R2), and Region-3 (R3) (see text for region averaging). COSMIC derived water vapour is from 2-10 km, whereas SABER derived is from 15-65 km.

Figure 3. Same as Fig.2, but for diurnal mean removed.

Figure 4. Diurnal variability of temperature (diurnal mean removed) for different seasons of Region-1 (R1), Region-2 (R2), and Region-3 (R3). COSMIC derived temperature is from 0-35 km, whereas SABER derived is from 15-65 km.

Figure 5. Diurnal variability of water vapour mixing ratio (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 Region-1 (R1), Region-2 (R2), and Region-3 (R3).

Figure 6. Diurnal variability of mean removed water vapour mixing ratio (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 Region-1 (R1), Region-2 (R2), and Region-3 (R3). Cold-point tropopause altitude (CPT-A) and temperature (CPT-T) are estimated for Region-1 and -2, whereas lapse rate tropopause (LRT) is estimated for Region-3, as well-defined CPT cannot be obtained for Region-3, except for JJA. Vertical lines indicate the respective standard errors.

Figure 7. Height profiles of correlation coefficient (r) between WMR and temperature (T) for different seasons of Region-1, Region-2, and Region-3. The profiles from 5-10 km (5-25 km) are from COSMIC derived WMR (T), whereas the profile from 15-60 km is from SABER derived WMR and T.

Figure 8. Height profiles of diurnal amplitude and phase of (a) WMR, and (b) temperature for different seasons at Region -1, -2, -3. The profiles from 5-10 km (5-25 km) are from COSMIC derived WMR (T), whereas the profile from 15-60 km is from SABER derived WMR and T.

Supplementary Figures

Figure S1. Temperature profiles (black) before (left) and after (right) quality control for January (2007-2019) between 0-10°N and 60-100°E. Red and green lines in left panel indicate the -3σ and $+3\sigma$, respectively for the month of January of same grid.

Figure S2. Water vapour mixing ratio (WMR) profiles (black) before (left) and after (right) quality control for January (2007-2019) between 0-10°N and 60-100°E. Red and green lines in left panel indicate the 0.25 percentile and 99.75 percentile, respectively for the month of January of same grid.

Figure S3. Same as Fig.2, but for standard errors (water vapour).

516 **Figure S4.** (left panels) Time-series of water vapour mixing ratio for different seasons at (a) 6
517 km, (b) 8 km, (c) 16 km, (d) 25 km, (d) 40 km, and (e) 50 km along with its (right panels)
518 standard error for Region-1 (R1), Region-2 (R2), and Region-3 (R3).

519 **Figure S5.** Same as Fig.4, but for standard error (temperature).

520

521

Figure 1. Seasonal mean latitude-longitude distribution of water vapour at 6, 8, 16, 25, 40, and 50 km for December-January-February (DJF), March-April-May (MAM), June-July-August (JJA), and September-October-November (SON). Water vapour mixing ratio up to 8 km is measured using COSMIC-1 from December 2006 to November 2019, and from 16 km onwards to higher levels are derived from TIMED-SABER from January 2002 to December 2019. Black closed contour (at 6, 8 and 16.5 km) indicates the climatological location of the Asian summer monsoon (ASM) anticyclone during JJA as defined by Park et al., (2007) by defining a stream function contour.(400 $\text{m}^2 \text{s}^{-1}$) (for more detail please refer Park et al. 2007 and Santee et al. 2017).

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(e)

(f)

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