Spatial and temporal variation of water vapour in upper troposphere and lower stratosphere over Indian region

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Spatial and temporal distribution of water vapour in the upper troposphere and lower stratosphere (UTLS) region over India including Arabian Sea and Bay of Bengal is presented using COSMIC/FORMOSAT 3 radio occultation measurements. Water vapour plays a crucial role in many aspects of UTLS chemistry. The influence of Asian summer monsoon can be seen in the seasonal pattern of water vapour in the UTLS region. It is observed that water vapour in the lower stratosphere follows the seasonal cycle in upper tropospheric water vapour with a time lag of one month. The time scale of cross tropopause transport of air mass in the region is also discussed.

Keywords: Troposphere and stratosphere transport, upper troposphere and lower stratosphere, water vapour.

THE distribution and variability of water vapour in the upper troposphere and lower stratosphere (UTLS) region is important for a variety of scientific problems. Although there were some attempts in the past to summarize the distribution[1], the fundamental mechanism that controls distribution of water vapour in the UTLS region is still not clearly understood. Water vapour plays a prominent role in radiative balance of earth and it is a key tracer in UTLS region. Water vapour has important implications on global warming. Precise effects of changes in water vapour depend on altitude at which changes occur[2]. Water vapour is involved in several global warming feedback mechanisms. Currently even the net sign of the feedback is uncertain. Rising surface temperature should result in increase in water vapour because of increase in evaporation. The direct radiative effect of increased water vapour is positive global warming feedback. But it has been suggested[3] that increase in surface temperature will result in more vigorous convection that could tend to dry upper troposphere. This will cause negative global warming feedback. In addition to tropospheric warming, water vapour leads to radiative cooling in stratosphere. Current radiative problems require accurate knowledge of water vapour distributions in the UTLS region.

Water vapour distribution is influenced by atmospheric dynamics and also influences them in return. So water vapour distribution can be an indicator of atmospheric circulation systems. Water vapour is an important parameter in atmospheric chemistry. It is a major source of hydroxyl radical in both troposphere and stratosphere and plays a major role in atmospheric oxidation chemistry. Globally, lower stratospheric water vapour is influenced by various factors like meridional circulation, stratospheric chemistry, dehydration at tropical tropopause and tropospheric leakage to extratropical lowermost stratosphere through isoentropic surfaces. Globally lower stratosphere is dry and extremely dry over tropics. It is reported that water vapour in lower stratosphere shows seasonal and interannual variability[4]. Seasonal variability of water vapour is different in tropics, midlatitude and polar region. Seasonal variations highlight the importance of monsoon regions in transport of water vapour to lower stratosphere[5]. Asian monsoon region is the main source of upper tropospheric water vapour from May to September[6]. It is found that upper tropospheric water vapour shows maxima over south Asian and North American monsoon regions. This observation gives the importance of monsoon region, in air transport within UTLS region. In tropics, maxima is associated with ascending branches of monsoon systems. Upper tropospheric humidity shows minimum in subtropics, associated with descending branches circulation systems. Tropopause temperature has strong influence in controlling water vapour in lower stratosphere. However, this alone cannot explain observed characteristics of water vapour distribution. Reproduction of water vapour distribution and variability in lower stratosphere is a challenge in global climate models. In tropics, surface conditions are strongly coupled to upper troposphere, whereas in extratropics influence of surface conditions is less. The seasonally varying structures with spatial coherence between upper troposphere and lower stratosphere can be used as an evidence for stratosphere–troposphere exchange.

Upper troposphere is the transition region separating chemically different troposphere and stratosphere. The region exchanges air with the lower stratosphere and lower troposphere. The varying degree of efficiency of

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exchange processes occurring in the area determines effectiveness of photochemical processes. Understanding the physical basis of tropopause and its role in transport is central to accurate modelling of chemical processes occurring in the region. Seasonal variation of exchange process is found to be related to tropopause height[7]. The source of water vapour in stratosphere is methane oxidation and transport from troposphere. It is well agreed that irreversible troposphere to stratosphere air transport occur through tropics[8]. Different hypotheses are put forward to explain the cross-tropopause air movement in tropics. The stratospheric fountain hypothesis[9] suggests that cross-tropopause transport occurs through regions where tropopause temperature is showing zonal minimum under the influence of extratropical wave forcing. From stratospheric distribution of water vapour, it is estimated that cross-tropopause transport occurs through the regions where tropopause temperature is less than 191 K (ref. 10). Preferred locations of this transport were over western Pacific, Indonesia, Malaysia and over India during monsoons. According to this hypothesis, tropical convection transports air up to 12–13 km height; from there air movement is slow (at a rate 20–30 m/day) driven by large scale meridional circulation known as Dobson circulation[11], which depends on radiative heating, not on convection. Another hypothesis relaxed the necessity of air to enter stratosphere just above the minimum tropopause temperature[12]. The horizontal transport may cause air experienced exceptionally low temperature to enter stratosphere at locations far away from observed minimum tropopause temperature. Another mechanism is rapid overshooting convection, which brings tropospheric air directly into stratosphere[13]. But the relative importance of these two mechanisms in controlling troposphere stratosphere exchange and hence lower stratospheric water vapour is not clear. Levine *et al.* reported that troposphere to stratosphere transport in tropical region is dominated by transport into extratropical lowermost stratosphere through isoentropic surfaces[14].

Time scale of transport from troposphere to stratosphere is important for water vapour and many halogenated compounds having very short life time[15]. Tropics are frequently encountered by convective activities. Convection can bring air parcels from boundary layer to the base of tropical tropopause layer (TTL) very rapidly. The movement of air parcel inside TTL is slow except in cases of overshooting convection. The stratospheric entry of these compounds depends on the time scale of transport within this layer comparable to life time of these compounds.

Spatial and temporal variation of water vapour in UTLS over tropical Indian region in the lat. 0–30°N and long. 40–120°E for one year period from October 2006 to September 2007 deduced from radio occultation measurements is presented here. The concentration and annual cycle of water vapour in both the regions is used to understand the cross-tropopause transport time scale.

Database

The source of data for this study is radio occultation observations from Constellation Observing System for Meteorology Ionosphere and Climate (COSMIC)/ Formosa Satellite 3 (FORMOSAT-3) satellite mission. The radio occultation technique has the advantage of global coverage, high accuracy, high vertical resolution and is free from weather constraints[16]. The mission was launched in April 2006 and presently makes use of six low earth orbit (LEO) satellites each of which give high-vertical resolution global observations of atmospheric parameters as GPS signals are getting occulted by earth limb. During initial months, daily occultations were less and increased enormously from August 2006. Technical and operational details of COSMIC mission were described earlier[17,18]. The accuracy of COSMIC mission is verified with previous radio occultation measurements CHAMP and GRACE[18]. COSMIC mission provides observations at 0.1 km altitude interval.

The GPS radio occultation technique is useful in collecting global high-resolution atmospheric temperature and water vapour data. Comprehensive validation studies of radio occultation measurements were reported. Comparison of vertical profiles of specific humidity, temperature and refractivity from COSMIC measurements with radiosonde, CHAMP, ECMWF, NCEP observations in the latitude range (20°N–20°S) showed good agreement above 4 km (ref. 19). Rao *et al.*[20] compared COSMIC water vapour profiles with high-resolution radiosonde measurements over a tropical site Gadanki (13.48°N, 79.2°E) in wet region of atmosphere. Both measurements showed good agreement with a mean difference of 5–10% below 6–7 km. They also performed comparison of temperature measurements in UTLS and found that bias between both measurements is less than 1 K in the altitude range 10–27 km (ref. 20). Comparison of water vapour profiles from CHAMP with radiosonde measurements showed a bias of less than 0.1 g/kg and standard deviation less than 1 g/kg in mid troposphere[21]. Comparison of COSMIC temperature observations with radiosonde measurements over Ausralia showed good agreement with a difference of average temperature ~0.5°C between them[22].

Data analysis

Tropopause is identified using cold point criterion. In tropical region, cold point tropopause criteria is more suited as the region frequently encounters convective activities. The criterion 3 km below tropopause is used for upper troposphere and 3 km above tropopause is used for lower stratosphere. In tropics, mean tropopause height is found at ~17 km. The criterion 3 km below and above tropopause is taken as it corresponds to the level of zero net radiative heating ~14 km and the level of stable stratospheric lapse rate ~20 km respectively. Average water vapour concentration in each profile is estimated for 3 km above tropopause and 3 km below tropopause. Monthly mean of water vapour and temperature in UTLS region is estimated using $10° \times 10°$ grid interval for the lat. 0–30°N and long. 40–120°E.

Results

Upper tropospheric water vapour

From the data analysis, it is seen that the monthly mean of water vapour in upper troposphere is found to vary between 4.3 and 9.6 ppm. Figure 1 shows monthly distribution of water vapour in UTLS averaged over whole range under study. It is found that monthly mean of water vapour shows seasonal pattern with minimum in the period from January to April. It starts increasing from May onwards and reaches maximum value in August and then starts decreasing. This increase in water vapour can be associated with Asian summer monsoon and strong convective activities persisting in the region during that period. Several workers have reported the influence of Asian summer monsoon on upper tropospheric water vapour[23–25]. Our results are in agreement with previous studies. It is found that upper tropospheric water vapour and atmospheric circulation systems are closely related[26]. The maximum of upper tropospheric water vapour is found to occur with upper level divergence and minimum with upper level convergence in circulations[26]. It is observed that upper troposphere divergence existing over Asian monsoon region during June–August and upper level convergence in December–February[26]. The pattern of water vapour variation in our observations is in conformance with the circulation patterns existing in the region.

The variation of water vapour in zonal direction is done by averaging over 0–30° latitude range at 10° longitude interval. Zonal variation in distribution of water vapour is shown in Figure 2. It is seen that the upper tropospheric water vapour is less in the 40–60 longitude range which covers more of the Arabian Sea region than the Bay of Bengal and Indian land mass. We investigated zonal distribution separately for all months; it is found that there is difference in longitudinal distribution except in February. During summer monsoon period, SST over the Arabian Sea region is found to be less than that in the Bay of Bengal region[27]. So atmospheric convective activities are less over the Arabian Sea region than that over the Bay of Bengal and Indian land mass. The difference in convective activities between two regions will be the reason for longitudinal variation in upper tropospheric water vapour. The meridional variation in water vapour averages over 40–120°E longitude range at 10° latitude interval. Meridional distribution of water vapour in upper troposphere is shown in Figure 3. Meridionally water vapour in upper troposphere decreases with latitude. Observations are in accordance with earlier studies, which reported upper troposphere water vapour minima in subtropics[28]. Maxima and minima in upper tropospheric water vapour are associated with ascending and descending branch of meridional circulation respectively. Apart from deep convection, spatial distribution of water vapour in the UTLS region is found to depend on monsoon anticyclone and slow ascent through TTL[29]. According to them, the highest values of water vapour in Asian monsoon region exist over the areas of anticyclone[29,30], which are located on north-west side of regions of deep convection[31,32]. The region of monsoon anticyclone is found to exist over Iran, Northern India, Tibet and China[31], whereas strongest convection occurs over the

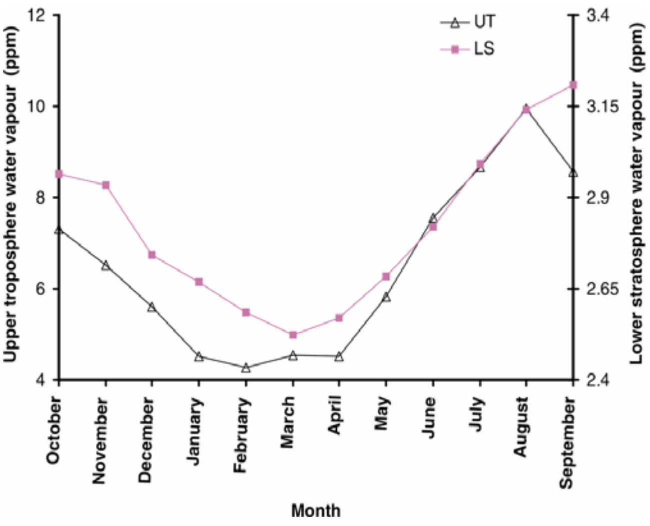

Figure 1. Monthly mean of water vapour in upper troposphere and lower stratosphere (UTLS) averaged over 0–30°N and 40–120°E.

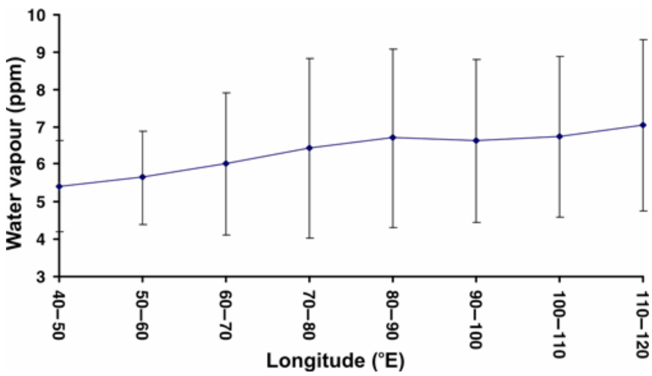

Figure 2. Zonal distribution of water vapour in upper troposphere (vertical bars indicate standard deviation of monthly mean values).

Bay of Bengal and South East Asia. But we found that during monsoon also water vapour concentration is more in the regions of deep convection than in the regions of anticyclone.

Lower stratospheric water vapour

Monthly mean of water vapour in the lower stratosphere region is found to vary between 2.5 and 3.2 ppm. Stratospheric water vapour also shows minimum from January to April and then slowly increases reaching maximum in September and then starts decreasing. Observations are in agreement with previous studies, which reported similar pattern with maximum in August–November period and minimum in January–April in tropical lower stratosphere[28]. It is reported that in northern hemisphere, lower stratospheric water vapour shows minima in winter and maxima in summer[33]. Entry of air from troposphere to stratosphere is influenced by south Asian summer monsoon[6]. The seasonal variation of lower stratospheric water vapour can be attributed to the influence of south Asian summer monsoon and atmospheric circulation systems. The variation of water vapour in lower stratosphere is also found to follow the pattern observed in upper troposphere. Meridional distribution of water vapour for different months is shown in Figure 4. Although upper tropospheric water vapour shows meridional and zonal variations, no such pattern is observed in lower stratosphere. Zonal distribution of water vapour averaged over 0–30 latitude at 10° longitude interval is shown in Figure 5. Here we are considering air movement to stratospheric overworld only. Levine *et al.* show that water vapour entry into stratospheric overworld exhibits less longitudinal structure[14]. Spatial distribution of water vapour in lower stratosphere indicates that air mass in the UTLS region undergoes sufficient horizontal mixing processes. Few studies suggest the horizontal movement of air in tropical UTLS region[28]. It is also reported that spatial variation in water vapour pattern at tropopause level vanishes in lower stratosphere[35,36]. It is suggested that the relative short time scale of horizontal motion relative to vertical transport near tropopause region may be the reason for the observed feature[12]. Rapid zonal and meridional mixing of ascending air masses is reported in the vicinity of tropopause[36].

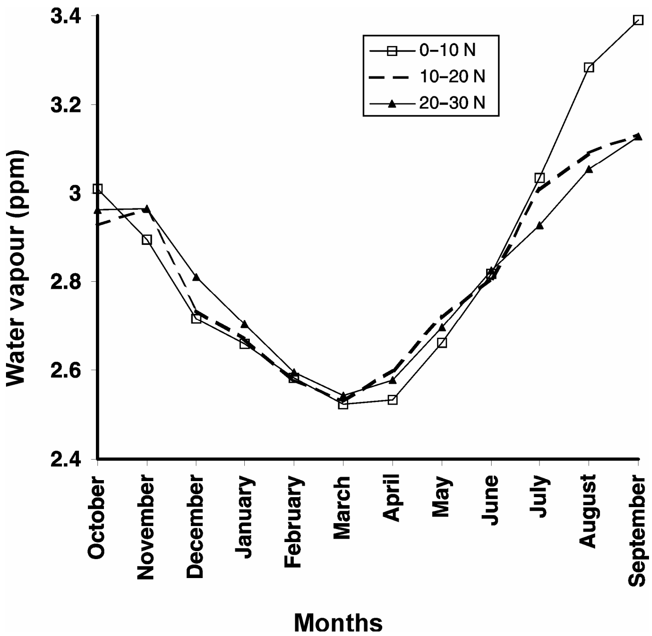

Figure 4. Monthly distribution of water vapour in lower stratosphere for different latitude range averaged over 40–120° longitude.

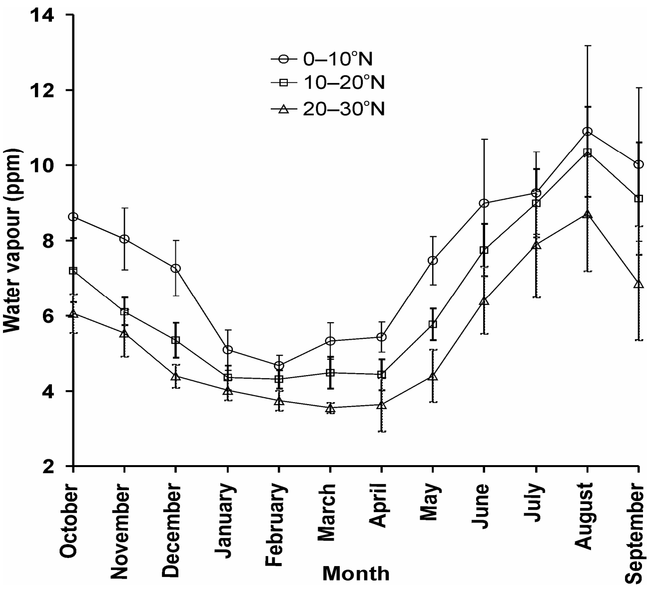

Figure 3. Monthly variation of upper tropospheric water vapour for different latitude range (vertical bars indicate standard deviation of zonal mean values).

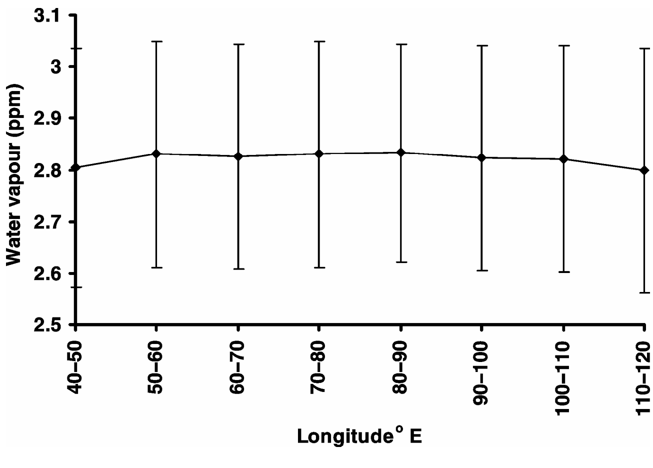

Figure 5. Zonal distribution of lower stratospheric water vapour averaged over 0–30°N latitude (vertical bars indicate standard deviation of monthly mean values).

Troposphere stratosphere transport

It is observed that monthly mean of water vapour in the upper troposphere shows maximum in August and minimum in February, though the variation is small from January to March (~0.3 ppm). In lower stratosphere maximum is observed in September and minimum in March. Water vapour minima and maxima in lower stratosphere lags by one month with respect to corresponding observations in the upper troposphere. It can be considered that air mass in upper troposphere takes one month to reach lower stratosphere. Further, we estimated 15 day average of water vapour profiles to find out whether time lag in appearance of maxima and minima is less than one month. Semimonthly distribution of water vapour in upper troposphere and lower stratosphere is presented in Figure 6. The distribution pattern is not smooth and uniform as observed in the case of monthly mean. Small fluctuations from seasonal pattern can be seen in some months. It is found that a time lag of 15 days occurs in the case of appearance of maximum and one month in the case of minimum between UTLS respectively. Smith *et al.*, using HALOE measurements, reported that the influence of water vapour increase in tropics at 105 mb (~17 km) in August–September period appears at 70 mb pressure level (~19 km) in January[23]. So it takes four months time for air mass to travel from tropical upper troposphere to lower stratosphere. This time lag is in agreement with the speed of Dobson circulation. On the basis of atmospheric general circulation model together with satellite observations[35] it is reported that in winter, water vapour contour at 100 mb altitude level appears at 75 mb in one month time and in summer it takes 2–3 months. The difference in extra tropical wave forcing is attributed to this seasonal variability. Similar time lag of three months for water vapour transport from 100 mb level to 80–90 mb is observed in HALOE measurements[36]. Our observations show that air mass movement in the UTLS region is comparatively faster than that reported by previous investigators. The area of our study coming under South Asian monsoon region frequently encounters deep convective activities[37]. Overshooting convections are too rapid to explain in terms of observed time lag. Using tracer studies it is found that in extra tropics, cross-tropopause transport from lower most stratosphere to troposphere takes one month time[38]. In extra tropics movement from lowermost stratosphere to troposphere is found to be controlled by small scale activities like tropopause folds and anticyclones. It is found that in tropics, troposphere to stratosphere transport is enhanced by mechanisms like cloud lofting, gravity waves and Kelvin waves[39–41]. The time lag of one month in crossing the tropopause in our studies may be due to modification of large scale ascent by small scale processes like cloud lofting, gravity waves and Kelvin waves in the region.

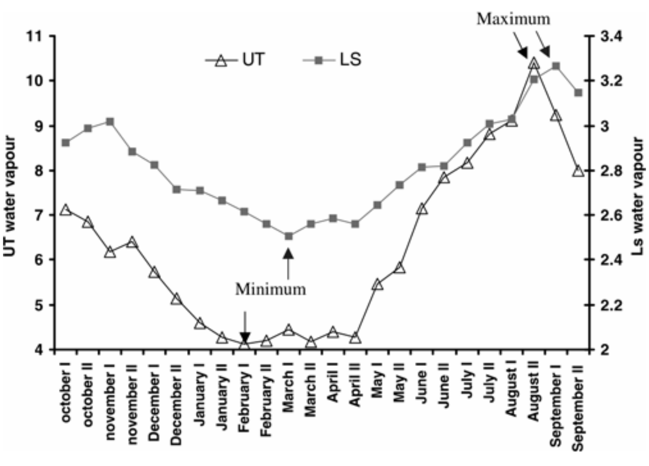

Figure 6. Semimonthly variation of water vapour in UTLS (UT stands for upper troposphere and LS for lower stratosphere. Maximum and minimum values are indicated by arrows).

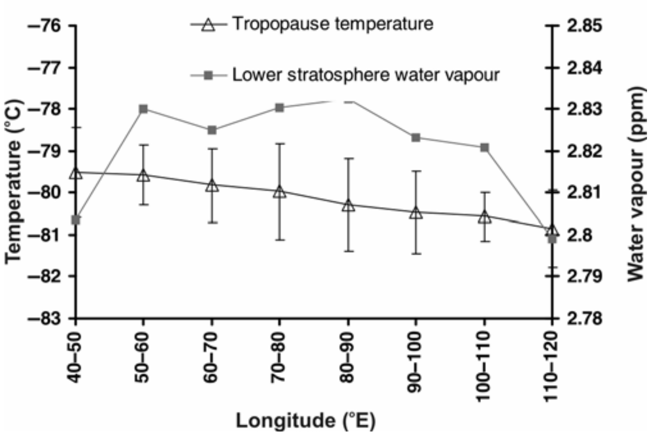

Figure 7. Zonal distribution of tropopause temperature and lower stratospheric water vapour averaged over 0–30°N (vertical bars indicate standard deviation of monthly mean values).

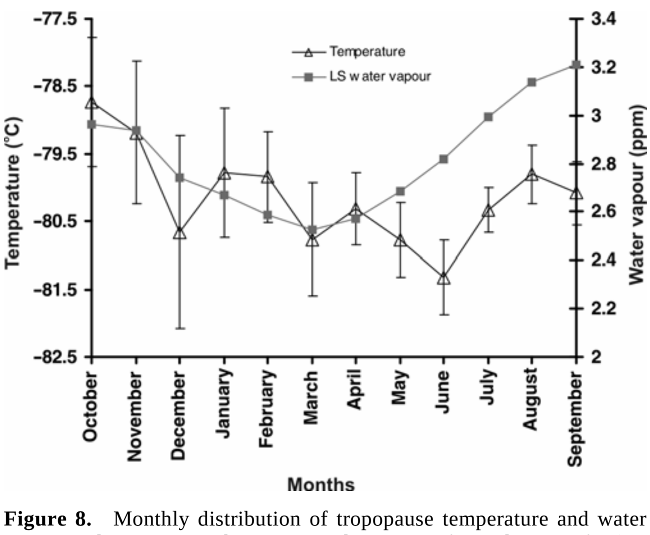

Figure 8. Monthly distribution of tropopause temperature and water vapour in lower stratosphere averaged over 0–30°N and 40–120°E (LS stands for lower stratosphere, Verical bars indicate standard deviation in temperature observations).

It is well established that tropopause temperature has strong influence on water vapour entering stratosphere. Influence of tropopause temperature on water vapour concentration in lower stratosphere is studied. Zonal variation of tropopause temperature and lower stratosphere water vapour is shown in Figure 7. Monthly distribution of tropopause temperature and lower stratosphere water vapour is shown in Figure 8. It can be seen that one to one correspondence does not exist between tropopause temperature and lower stratospheric water vapour. It may be due to some localized mechanism modulating the influence of tropopause temperature. Mean monthly distribution of tropopause temperature is found to exhibit some intraseasonal oscillations.

Conclusion

Water vapour distribution in UTLS region shows seasonal variation. It may be due to influence of south Asian summer monsoon on UTLS region. Influence of geographical region and surface conditions can be seen clearly in upper troposphere water vapour, but less significant in lower stratosphere. It is seen that signatures of changes in upper tropospheric water vapour appear in lower stratospheric water vapour in one month. Thus, troposphere to stratosphere transport is faster than previously reported values and influence of small scale localized mechanism. The influence of horizontal mixing processes in UTLS region can be seen from spatial invariability of water vapour in lower stratosphere.

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