

Water vapor measurements at ALOMAR over a solar cycle compared with model calculations by LIMA

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[1] Microwave water vapor measurements between 40 and 80 km altitude over a solar cycle (1996–2006) were carried out in high latitudes at Arctic Lidar Observatory for Middle Atmosphere Research (ALOMAR) (69.29°N, 16.03°E), Norway. Some smaller gaps and three interruptions of monitoring in the winters 1996/1997 and 2005/2006 and from spring 2001 to spring 2002 occurred during this period. The observations show a distinct year-to-year variability not directly related to solar Lyman- α radiation. In winter the water vapor mixing ratios in the upper domain were anticorrelated to the solar activity, whereas in summer, minima occurred in the years after the solar maximum in 2000/2001. In winter, sudden stratospheric warmings (SSWs) modulated the water vapor mixing ratios. Within the stratopause region a middle atmospheric water vapor maximum was observed, which results from the methane oxidation and is a regular feature there. The altitude of the maximum increased by approximately 5 km as summer approached. The largest mixing ratios were monitored in autumn. During the summer season a secondary water vapor maximum also occurred above 65 km most pronounced in late summer. The solar Lyman- α radiation impacts the water vapor mixing ratio particularly in winter above 65 km. In summer the correlation is positive below 70 km. The correlation is also positive in the lower mesosphere/stratopause region in winter due to the action of sudden stratospheric warmings, which occur more frequently under the condition of high solar activity and the enhancing the humidity. A strong day-to-day variability connected with planetary wave activity was found throughout the entire year. Model calculations by means of Leibniz-Institute Middle Atmosphere model (LIMA) reflect the essential patterns of the water vapor variation, but the results also show differences from the observations, indicating that exchange processes between the troposphere and stratosphere not modeled by LIMA could have influenced the long-term variability. We show results of measurements, compare these with calculations, and discuss the chemical and dynamical backgrounds of the variation of water vapor in the middle atmosphere.

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1. Introduction

[2] In the mesosphere, water vapor is the most important minor constituent, and in the whole middle atmosphere this is the case along with ozone. The reason is chiefly that water vapor is the main source of the chemically active hydrogen radicals which affect the chemistry of all other chemically active minor constituents. Water vapor itself plays a role in

various atmospheric phenomena, such as the creation of ice particles responsible for the so-called polar mesospheric summer echoes (PMSEs) and the formation of noctilucent clouds (NLCs). Moreover, it determines the production of water cluster ions in the mesopause region and it influences the thermal regime of the atmosphere. Considering both loss and production, its effective lifetime in the middle atmosphere is very long, amounting from several months up to infinity. Below about 65 km the lifetime is even negative, meaning that water vapor is not decomposed but is formed from the source species methane and molecular hydrogen [Sonnemann *et al.*, 2005]. Only in and above the upper mesosphere, water vapor is destroyed and converted into molecular or atomic hydrogen. The lifetime in the lower

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thermosphere amounts to more than a week. The most important photolyzer is the solar Lyman- α radiation which varies by nearly a factor of 2 from solar minimum to maximum [Woods *et al.*, 2000]. The lifetime of water vapor is on the order of transport time scales even in the lower thermosphere [Stevens *et al.*, 2003] and can be used as tracer for dynamical processes such as planetary waves [Sonnemann *et al.*, 2008].

[3] The anthropogenic growing methane concentration in the troposphere gave rise to the supposition that the water vapor concentration in the middle atmosphere had also increased in the past [World Meteorological Organization, 1999; Khalil *et al.*, 1993; Dlugokencky *et al.*, 2003] and that even the occurrence rate of NLCs could be influenced by the rising methane concentration [Thomas *et al.*, 1989; Thomas and Olivero, 2001]. Long-term trend calculations confirmed a considerable increase of the mesospheric water vapor mixing ratio [Grygalashvily and Sonnemann, 2006; Grygalashvily *et al.*, 2010]. However, in the recent past this trend seems to have stopped or slowed down considerably [Khalil *et al.*, 1993; Dlugokencky *et al.*, 2003]. It was stated that only a certain portion of the observed increase of the stratospheric humidity in the recent past can be attributed to the increase of methane and that the particular exchange conditions of water vapor itself between the troposphere and stratosphere essentially influence the middle atmospheric humidity [de F. Forster and Shine, 1999]. The general mechanism of the water vapor exchange is an upward transport in the tropical convection zone. At the tropical tropopause layer (TTL) it is subjected to the freeze drying effect. The water vapor mixing ratio decreases to about 4 ppmv there. Above the hygropause the mixing ratio of water vapor increases with increasing height due to the oxidation of methane and reaches values of 7 to 8 ppmv at the stratopause. In middle and high latitudes, stratospheric air enters the troposphere, conveying more humid air into this domain than opposite coming from the troposphere in the tropics into the stratosphere [Sonnemann and Körner, 2003]. The injection of stratospheric air containing stratospheric water vapor as well as ozone is forced by the impact of low-pressure troughs [Junge, 1962; Johnson and Viezee, 1981], and this process takes place particularly in the winter polar vortex. As the mesospheric water vapor is controlled by the stratospheric water vapor concentration, its long-term monitoring also provides information about the entire middle atmosphere, especially with regard to its long-range transport and exchange processes between the different atmospheric layers.

[4] A comparison between observations and model calculations helps understanding the observed variations and gives us deeper insight into the physical-chemical processes. A model can be complex and can provide information about quantities which are impossible to measure, such as further constituents and their spatiotemporal variability, or it can fill in gaps in a measuring series. Differences between observed measurements and model calculations give rise to investigating the cause of the discrepancy: are the measurements accurate or are the calculations wrong? Water vapor measurements in high latitudes, particularly in summer over nearly a solar cycle, are of particular interest in order to understand the formation and function of NLCs and PMSEs. Different indications point to an influence of the solar

activity upon the formation of these phenomena [e.g., Gadsden, 1998; Thomas and Olivero, 2001; DeLand *et al.*, 2003; von Zahn *et al.*, 2004], but there are still many open questions about the real processes forming ice particles in the summer mesopause region. Measurements at a fixed location have the advantage of reducing different variable impacts and enable us to investigate temporal variations at the same place. Ground-based microwave measurements are a particularly suitable tool for this. Some groups have measured water vapor in this way over longer periods [e.g., Bevilacqua *et al.*, 1985; Neduluha *et al.*, 1996], but they did not measure this constituent at high latitudes.

[5] In section 2 we briefly introduce and discuss the microwave facility and the GCM Leibniz-Institute Middle Atmosphere model (LIMA). In the presentation of results we present only analyses of the general trend in the period of monitoring and consider the possible dependence of the water vapor mixing ratio on the solar activity represented by the solar Lyman- α radiation. In the discussion we analyze the global dynamical connections entailing a time variation of mesospheric water vapor, and we debate possible reasons of some discrepancies between the observations and the calculations. Finally the main findings are summarized.

2. Description of the Microwave Instrument and the Model LIMA

[6] The microwave technique is used to investigate the composition of the middle atmosphere with respect to certain measurable minor constituents such as water vapor or ozone. The microwave facility at Arctic Lidar Observatory for Middle Atmosphere Research (ALOMAR) (69.29°N, 16.03°E), Norway, was described in detail by Hartogh and Hartmann [1990], Hartogh and Jarchow [1995], Seele and Hartogh [1999] and by Hartogh *et al.* [2004], among others. A ground-based heterodyne receiver detects the rotational transition of water vapor at 22.235 GHz. A steep single sideband waveguide filter selects the water line in the lower sideband. This filtered signal is multiplied with a 22.535 GHz local oscillator signal using a Schottky mixer. The 22.235 GHz signal is down-converted to an IF (intermediate frequency) of 300 MHz and fed into a Chirp Transform Spectrometer with a 40 MHz bandwidth and 20 KHz spectral resolution. Between 2002 and 2005 a second Chirp Transform spectrometer with about 200 MHz bandwidth and 50 kHz spectral resolution was operated [Villanueva and Hartogh, 2004; Villanueva *et al.*, 2006]. The effect of adding bandwidth can be seen in Figure 1a. Whereas before 2002 the water vapor profiles below 40 km represented the a priori profile used for the retrieval, from 2005 on the effect of the larger bandwidth in this altitude regime can clearly be recognized.

[7] The vertical resolution of the water vapor retrievals depends on the height and signal-to-noise ratio of the measurements. It varies between about 7 and 10 km (derived from the full width half maximum or the averaging kernels) and the integration time amounted to 1 day, which was necessary to reduce the error in the upper domain. The upper border of monitoring has a focal point at about 80 km, denoted as 80 km panel. Here the true water vapor profile experiences the strongest smoothing, resulting in the larger vertical resolution. The same smoothing effect also applies

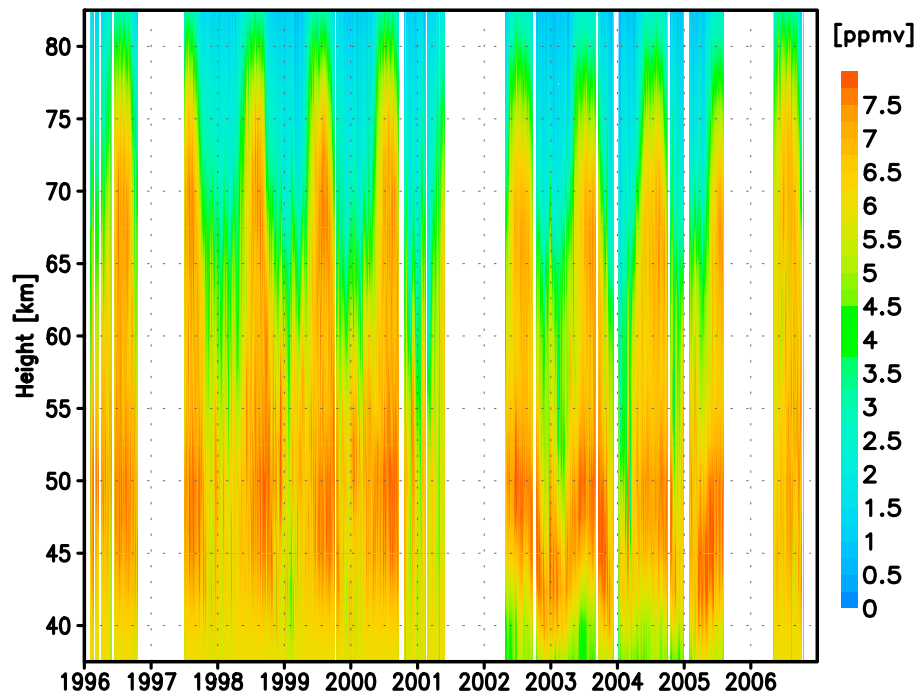


Figure 1a. Outline of the water vapor measurements at ALOMAR from the end of 1995 until the end 2006.

to the lower end of the vertical coverage. The respective altitudes appear to be about 40–45 km for the 40 MHz wide spectrometer and 25–30 km for the 200 MHz wide spectrometer. For both the high and the low end of the vertical coverage, the contribution of the a priori profile is still below 50%. Tropospheric conditions such as the temperature and humidity influence the accuracy of the measurements. The elevation angle of the observation is fixed to 18°, providing the best signal-to-noise ratio of the water line for the conditions at ALOMAR. The instrument observes southward, meaning the geographic position of the measurements in the upper mesosphere varies between about 68.04°N (at 80 km) and 68.67°N (at 40 km altitude) and does not cover 69.29°N, the latitude of ALOMAR. Depending on altitude, the estimated error amounts to 5 to 10%. For more details, see also *Seele and Hartogh* [1999]. Water vapor has been monitored at ALOMAR since the end of 1995. Three larger interruptions occurred in the winters 1996/1997 and 2005/2006 and from spring 2001 to spring 2002. Some smaller gaps were distributed over the time range of observations. For our analysis these gaps have been filled by linear interpolation.

[8] The abbreviation GCM LIMA stands for general circulation model Leibniz-Institute Middle Atmosphere. A comprehensive and detailed description of the dynamical part of the model was given by *Berger* [2008]. The chemistry transport model (CTM) of LIMA was introduced by *Sonnemann et al.* [2006, 2008]. The model LIMA is the successor of older versions of the model COMMA-IAP (Cologne Model of the Middle Atmosphere of the Institute of Atmospheric Physics) [e.g., *Sonnemann et al.*, 2005; *Hartogh et al.*, 2005; *Medvedev and Hartogh*, 2007, and references therein]. LIMA is a coupled model of dynamics

and chemistry. It is a fully nonlinear global three-dimensional Eulerian grid point model extending from 0 to approximately 150 km with a vertical resolution of about 1.1 km. The dynamic model has a new grid structure employing so-called reduced Gaussian coordinates. These coordinates are simplex or three-angle coordinates in the horizontal planes. The model possesses 41804 horizontal grid points entailing a mesh size of approximately 110 km. This small and nearly constant mesh size is advantageous for the considerations of the propagation of gravity waves and the avoidance of the so-called pole singularities. The main difference between LIMA and its predecessor COMMA-IAP is that COMMA-IAP was only able to calculate climatological averages whereas LIMA uses real tropospheric and lower stratospheric temperature and horizontal wind data up to 35 km height from assimilation of ECMWF/ERA-40 data (European Centre for Medium-Range Weather Forecast/Re-analysis Version 40). For the years prior to 2001 the model assimilates the so-called ERA-40 data set [*Uppala et al.*, 2005], and since 2002 we have been using the operational data set of ECMWF. The assimilated atmospheric data contains signals about planetary, gravity and tidal waves but also about the quasi-biennial oscillation (QBO) and further variations within the domain below 35 km. The waves can penetrate into the mesosphere/lower thermosphere (MLT region) and cause a pronounced internal variability in the upper atmosphere. In particular the model is also able to calculate the propagation of planetary waves and as a consequence it can numerically create sudden stratospheric warming events as shown by *Sonnemann et al.* [2006].

[9] The chemistry transport model takes into account three modules or codes: the chemistry, the transport and the

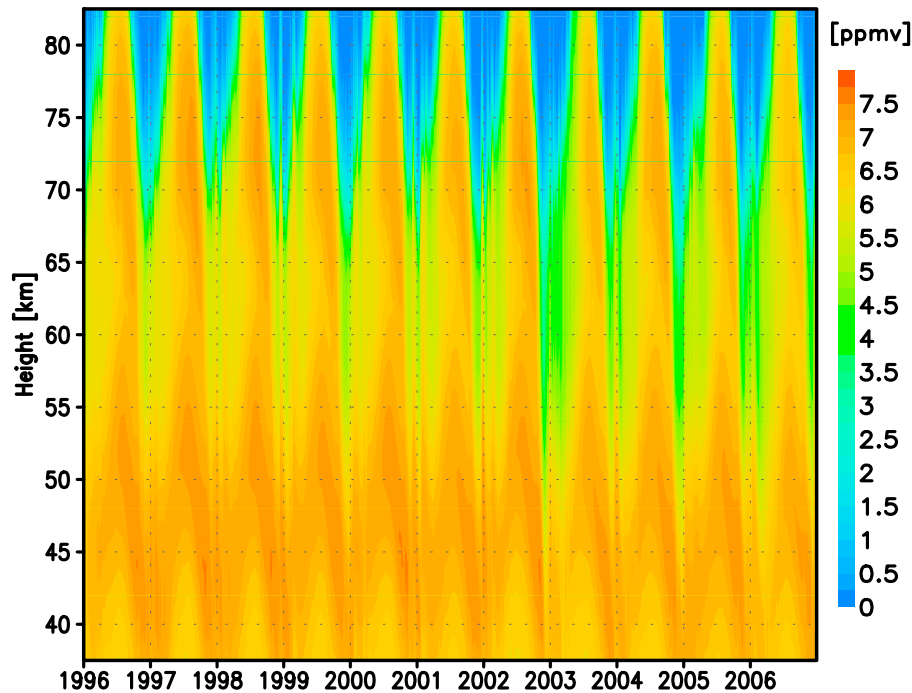


Figure 1b. Same state of affairs as shown in Figure 1a but for the LIMA calculations.

radiation module. The integration of the system is based on the operator splitting method. The chemical code takes into calculation all chemical species which are important for the chemistry of water vapor, meaning for the odd oxygen–odd hydrogen chemistry determining the chemistry of the middle atmosphere above 40 km [see *Körner and Sonnemann*, 2001; *Sonnemann and Körner*, 2003; *Sonnemann et al.*, 2005]. Long-lived hydrogen species are water vapor, molecular hydrogen, and methane. The odd oxygen constituents are atomic oxygen, $O(^1D)$, and ozone and the odd hydrogen compounds are the hydrogen radicals H , OH , HO_2 and H_2O_2 . The dynamical fields, temperature, pressure and the wind components, from the dynamical model are used in the CTM. The transport module takes into consideration the advective transport induced by the wind field and it considers both the turbulent and molecular diffusion. We have adopted eddy diffusion profiles for summer and winter of high latitudes according to results published by *Lübken* [1997] which do not depend on latitude and time. This is certainly a restriction influencing particularly the mesopause region but is fortunately less in the mesosphere itself. Spatiotemporal eddy diffusion profiles for use in chemical transport are not yet available. Additionally, we apply real data of the Lyman- α flux for the calculations according to *Woods et al.* [2000]. Data are available at: ftp://lasftp.colorado.edu/pub/SEE_Data/composite_lya/composite_lya.dat. A very important improvement of the model involves the implementation of a new transport scheme according to *Walcek* [2000] marked by an almost zero numerical diffusion. In order to integrate the system with a time step up to 1 min, the chemical module is based on a family concept [Shimazaki, 1985] as commonly used considering in the

mesosphere the odd hydrogen, the odd oxygen and the odd nitrogen (N , NO , NO_2 , NO_3) families.

3. Results

[10] Figure 1a shows an outline of the water vapor measurements at ALOMAR from the end of 1995 until the end 2006. Some smaller gaps and three interruptions of monitoring in the winters 1996/1997 and 2005/2006 and from spring 2001 to spring 2002 occurred during this period. Figures 1b and 1c display the same situation for the LIMA calculations. The annual variations exhibit the well-known patterns. The main water vapor maximum occurs at the stratopause region. The height of this peak varies by about 5 km over the year. The peak altitude is highest in August and occurs at about 50 km. Largest mixing ratios occur during autumn when the peak altitude has declined. A secondary maximum of the water vapor mixing ratio occurs at 65–70 km between the summer months and autumn, but it is absent during the rest of the year. The annual maximum in constant heights propagates downward from the upper mesosphere to the upper stratosphere beginning in late June through November. Particularly after winter solstice in the wake of sudden stratospheric warmings (SSWs) the water vapor mixing ratios increase so that the annual minima appear as early as late November/early December and a long-stretched second winter minimum occurs in the lower mesosphere in March/April. The stratospheric warming of February 1998 present in our data set has been analyzed in more detail. [Seele and Hartogh, 2000] illustrates the relationship of stratospheric warmings and the increase of water in the middle to upper stratosphere and mesosphere.

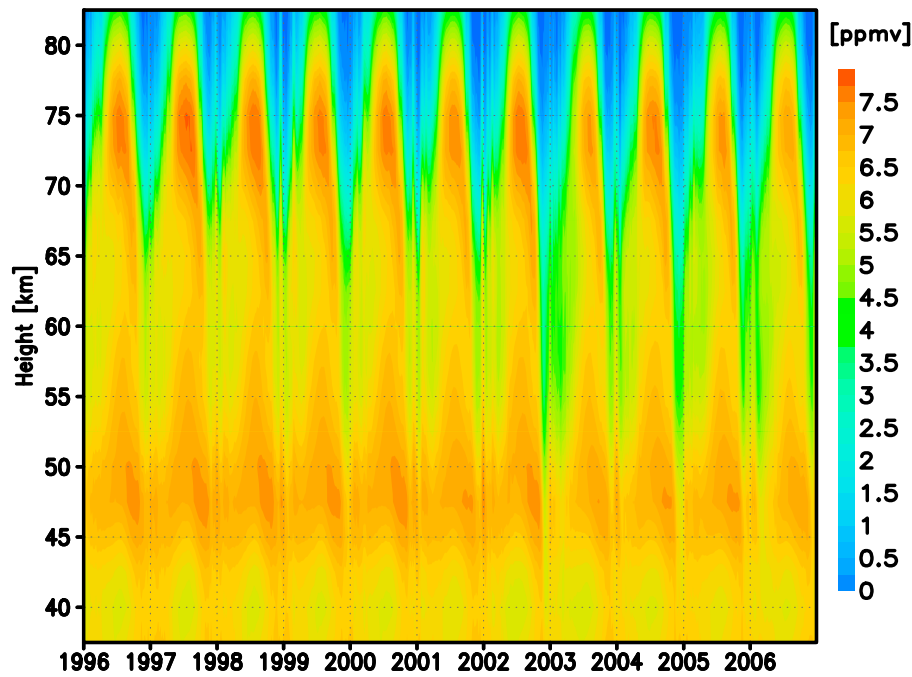


Figure 1c. Same data as used in Figure 1b but convolved with averaging kernels derived from the measurements.

[11] Figures 2a and 2b display the 7 day sliding average of the mean annual variation of the water vapor mixing ratio for the observations (Figure 2a, top), LIMA (Figure 2a, bottom) and LIMA results convolved with the averaging kernels of the observations (Figure 2b) and show the typical patterns and specifications. The largest differences between the observations and the calculations, even after convolution with the averaging kernels, occur in the upper mesosphere where in summer LIMA provides higher values which are comparable with HALOE measurements, but in winter the values are too small. A secondary water vapor maximum appears in Figures 2a and 2b but in the observations it lies between 65 and 70 km and in the calculations it is roughly 5 km higher, whereas the absolute values are very similar in the whole domain. The reason for the discrepancy between the observations and LIMA calculations could be an underestimation of the retrieved measured values in summer, a respective overestimation in winter at the uppermost panel, and/or a somewhat too strong vertical wind in the model in the upper mesopause. As mentioned before, the annual minimum values occur already in late November and early December before the SSW season starts. In the stratopause region a second slight long-stretched minimum occurs in late winter/early spring.

[12] Figure 3 displays the monthly mean values at 50, 60, 70, and 80 km altitude. Figure 3 demonstrates that the summer values at 70 km are equal to or often slightly higher than the values at 60 km. The most pronounced annual variation occurs at 70 km. This region is just the domain of the tertiary ozone maximum in winter and the low water vapor mixing ratios are one reason for the nighttime ozone

enhancement (the so-called tertiary ozone maximum) during this season.

[13] Figure 4 compares the ALOMAR observations with the LIMA calculations at 60 km showing the daily values. Generally, the observations are more variable than the calculations but the absolute values of the calculations fit the measurements pretty well. The agreement between observations and calculations improves after the 2001/2002 interruption. We speculate that both the assimilated data in the model (before 2001 the ERA-40 data set was used) and the microwave device, which was refurbished in 2002, became more exact after 2001/2002. The largest differences between observation and calculation occur at 80 km (not shown), where the model produces larger values in summer (comparable with HALOE data) and smaller values in winter compared to the observations.

[14] A cursory inspection of Figures 1a, 1b, 1c and 3 reveals an already decreasing tendency of water vapor mixing ratios. Particularly after 2001 the water vapor mixing ratios decreased suddenly, whereas during the years prior it seemed to increase slightly or to stagnate. This sudden decrease has been observed by other groups in low and mid latitudes [e.g., *Rosenlof and Reid, 2008*]. We present it for the first time in high latitudes. However, the behavior is different for the summer and winter months. Figures 5a and 5b exhibit trend analyses of the ALOMAR measurements and the LIMA calculations for the three summer months June, July and August (Figure 5a) and the three winter months December, January and February (Figure 5b) at the height levels 50, 60, 70, and 80 km. The results generally exhibit a decreasing behavior. This is particularly true for the LIMA calculations both for summer and winter months.

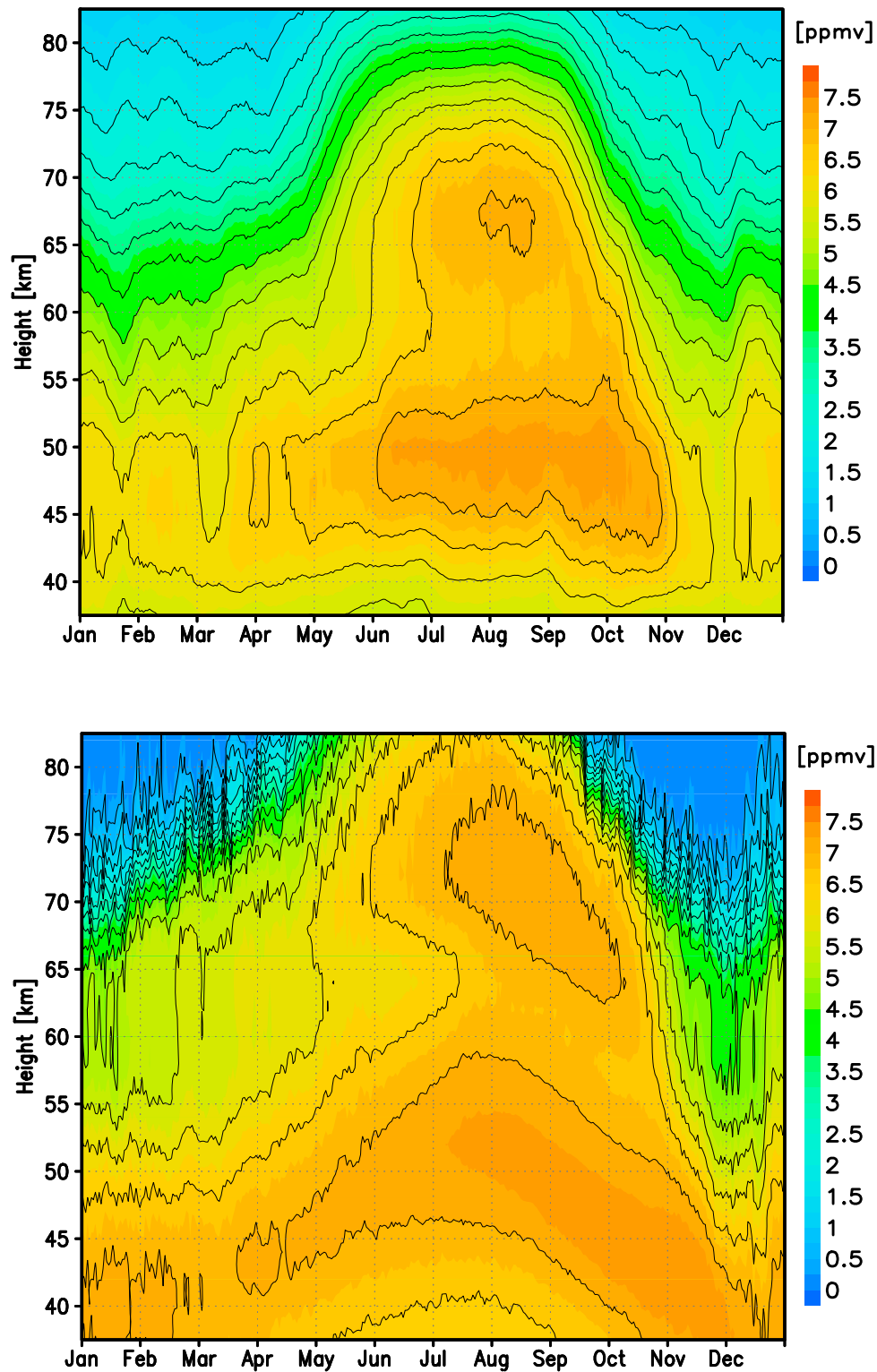


Figure 2a. Seven-day sliding mean of the annual variation of the water vapor mixing ratio averaged over 11 years for (top) the observations and (bottom) the calculations.

At 50 km in summer the correlation coefficient of the trend for the observations is very small with $r = -0.287$ and at 80 km a sudden decline occurred after the interruption of the monitoring in 2002. We discuss this finding in more detail

in section 4. In winter the behavior is essentially clearer, when a general tendency of water vapor decrease exists in the mesosphere. Because the monitoring interval which covers slightly more than a solar cycle begins shortly before

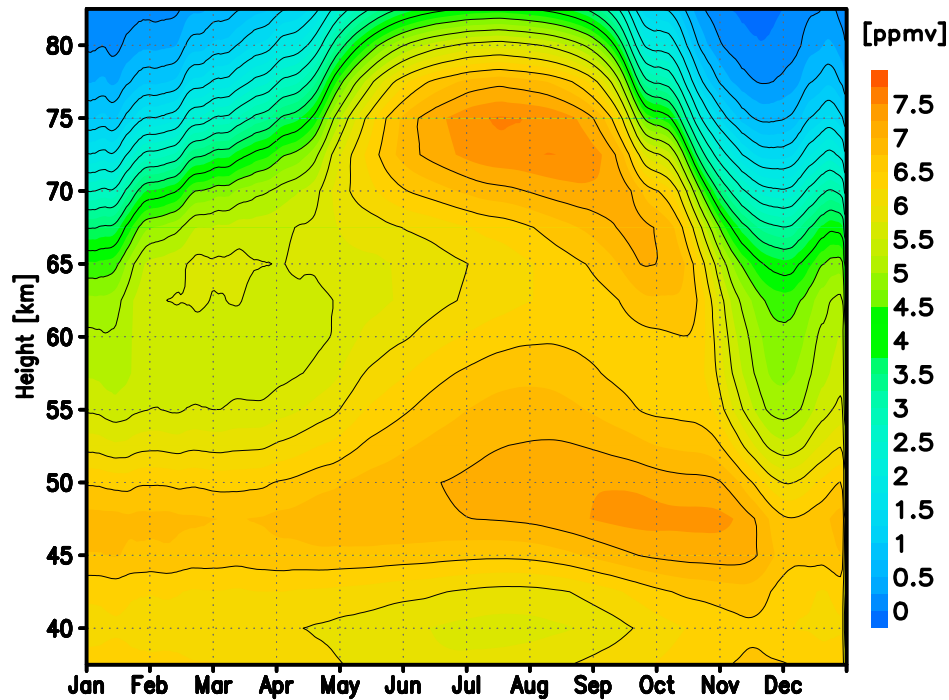


Figure 2b. Same data as used in Figure 2a (bottom) but convolved with averaging kernels derived from the measurements.

the solar activity minimum phase and ends also shortly before the next solar activity minimum phase (the maximum years were around 2001/2002), the declining tendency cannot directly result from the influence of the solar activity.

We note that the solar activity during the minimum phases does not considerably differ from one solar cycle to the next.

[15] Figures 6a and 6b depict the relative change of water vapor mixing ratio (in %) for an increase of the Lyman- α

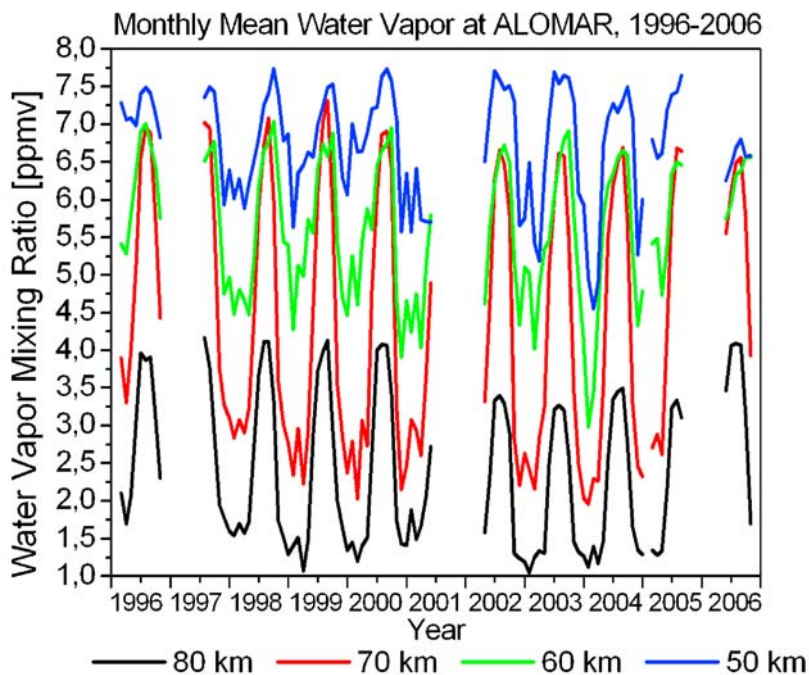


Figure 3. Monthly mean values of the water vapor mixing ratio measured at ALOMAR at 50, 60, 70, and 80 km altitude.

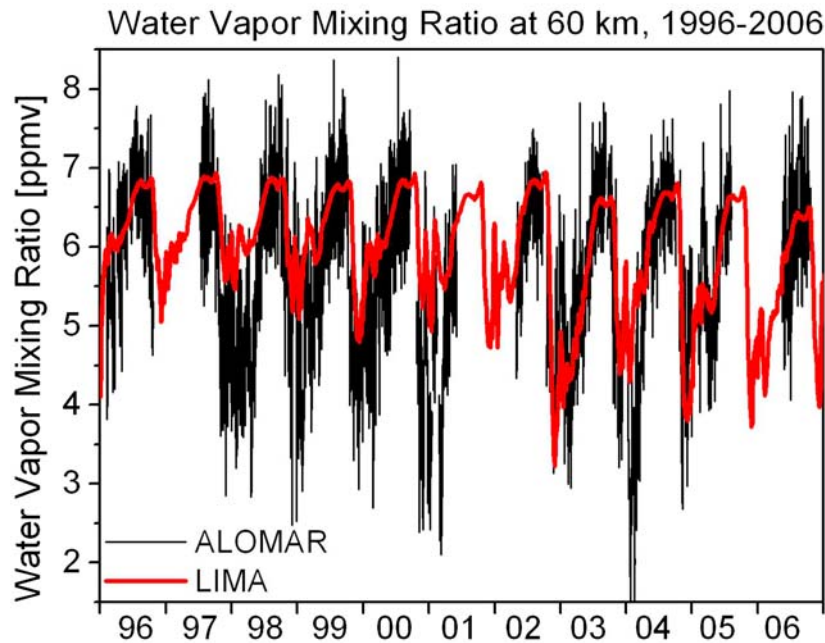


Figure 4. Comparison of the ALOMAR observations with the LIMA calculations at 60 km altitude.

radiation by 1×10^{11} photons $\text{cm}^{-2} \text{s}^{-1}$ for the ALOMAR observations (Figure 6a) and the LIMA calculations (Figure 6b) for summer and winter months. The lowest mean Lyman- α flux values did not fall below 3.5×10^{11} photons $\text{cm}^{-2} \text{s}^{-2}$. The uppermost value did not exceed 6.25×10^{11} photons $\text{cm}^{-2} \text{s}^{-2}$. As expected the winter values show a clear anticorrelation to the Lyman- α radiation in the middle and upper mesosphere with growing response with increasing height. In the lower mesosphere, however, the response is positive for LIMA or only weakly negative for the observations. One reason of this positive correlation in the lower mesosphere could be that SSWs occurred more frequently and were stronger during times of high solar activity [Sonnemann and Grygalashvily, 2007]. SSWs enhance the water vapor mixing ratio particular in the lower mesosphere [Seele and Hartogh, 2000]. This behavior is clearly demonstrated in Figures 1a–1c. The spikes in the water vapor mixing ratios in the lower mesosphere in winter result from the impact of SSWs.

[16] The response to the Lyman- α radiation is completely different in summer than in winter. The response is positive below 80 km for the calculated data and below 70 km for the observations with only a small negative value between 70 and 80 km. The reason seems to be understandable, as the optical depth of penetration of unity for the Lyman- α radiation ranges around 75 km depending on the solar zenith angle. The Lyman- α radiation is the most important radiation dissociating water vapor in the mesosphere which varies strongly with the solar activity by approximately a factor of 2 from minimum to maximum [Woods *et al.*, 2000]. During winter the downward directed vertical wind conveys air which is poor in water vapor and is impacted by the varying Lyman- α radiation from the mesopause region into the lower domain, whereas the situation is different in summer, when relatively humid air not strongly influenced

by the Lyman- α radiation is lifted upward. Below about 70–75 km the effective lifetime of water vapor is extremely large, on the order of several months, and even changes its sign below about 65 km [Sonnemann *et al.*, 2005]. As mentioned earlier, the effective lifetime includes both loss and production of the considered constituent. The largest part of dissociated water vapor returns to water vapor in some so-called zero cycles. But below about 65 km, water vapor is autocatalytically produced from the reservoir of molecular hydrogen and the rest of methane, thus increasing dissociating radiation amplifies this effect. This is the reason why, under conditions of high solar activity, water vapor increases in the domain below 65 km. However, the situation seems to be more complicated, as we discuss in section 4.

4. Discussion

[17] The negative trend of the water vapor mixing ratio is amazing because the anthropogenic growth of methane would be expected to increase the middle atmospheric humidity. (The word trend is not used in a climatological sense but as the tendency for a limited time interval of a solar cycle.) Methane is not subjected to the freeze-drying at the hygropause and can enter into the stratosphere in the tropics. However, on the one hand the methane increase has seemed to be stopped or slowed down in the recent past [Khalil *et al.*, 1993; Dlugokencky *et al.*, 2003], and on the other hand only one part of the stratospheric variation of water vapor can be contributed to the methane oxidation. The other part results from a natural variability [de F. Forster and Shine, 1999] such as the Brewer-Dobson circulation connected with exchange processes between troposphere and stratosphere. The global circulation also influences the water vapor transport in the mesosphere of high latitudes.

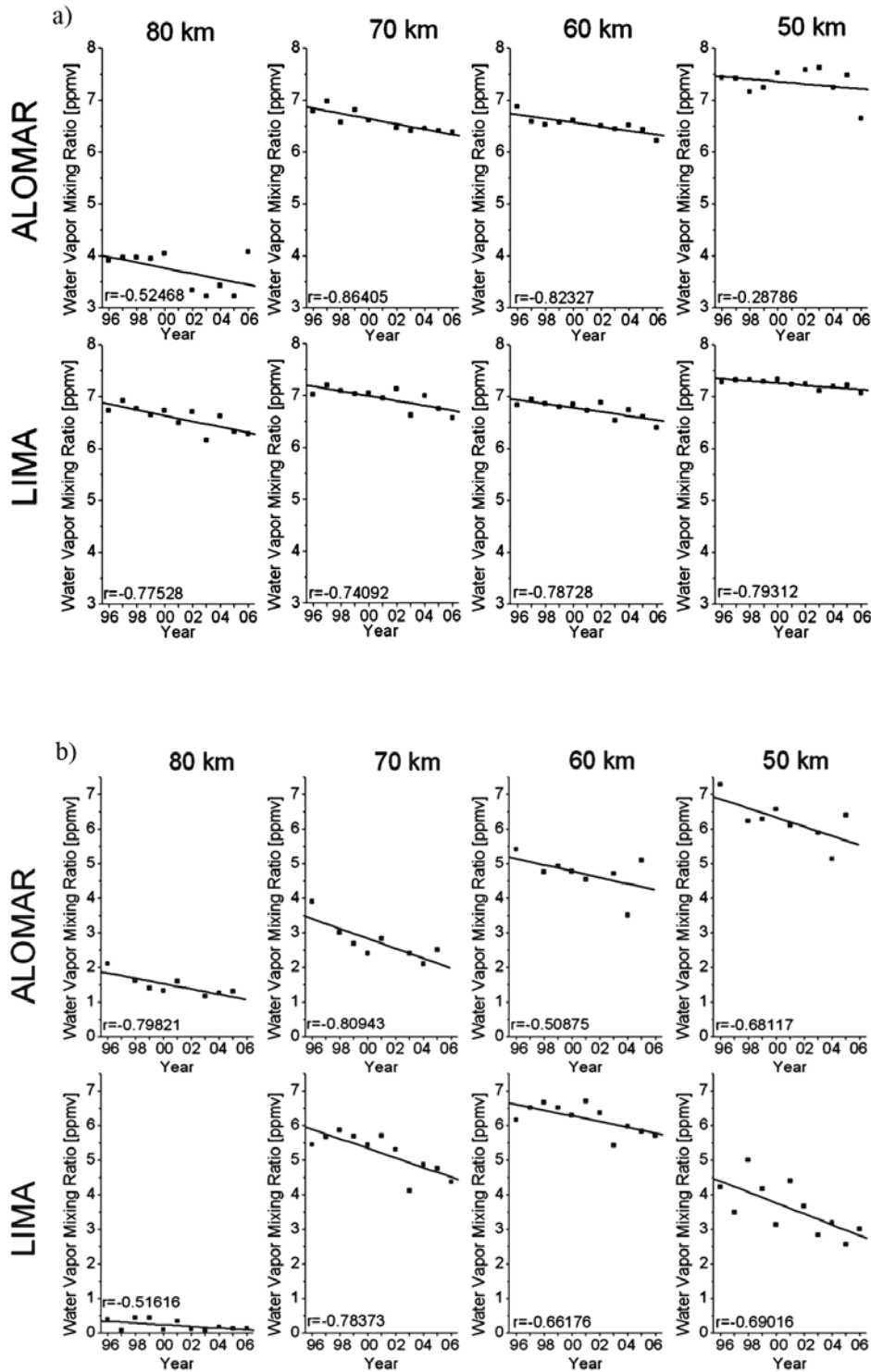


Figure 5. Correlation analyses of the ALOMAR measurements and the LIMA calculations for (a) the three summer months June, July, and August and (b) the three winter months December, January, and February at the height levels 50, 60, 70, and 80 km.

[18] LIMA does not consider the variable exchange processes at the hygropause. The chemical transport model uses identical lower boundary conditions at the hygropause which do not change from year to year. Despite this restriction it also calculates a decreasing trend of water vapor

in the mesosphere. As the model is only controlled by the assimilated data and the varying Lyman- α radiation, the internal middle atmospheric variability of the water vapor distribution results from the impact of the changing dynamics and radiation. The observations also contain, of

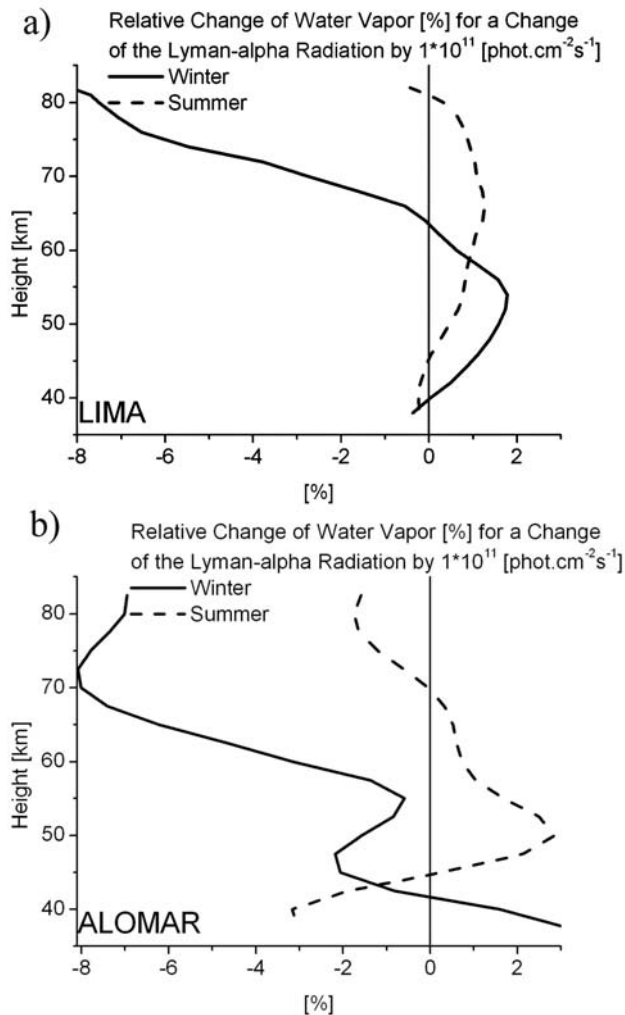


Figure 6. Relative change of water vapor mixing ratio (in percent) for an increase of the Lyman- α radiation by 1×10^{11} photons/cm² s⁻¹ for (a) the ALOMAR observations and (b) the LIMA calculations, with summer and winter months considered separately.

course, the variability resulting from the exchange processes. This fact and the interruptions of the monitored sequence, the second interruption just in the period of highest solar activity, may explain some differences between the observations and calculations as can be seen in Figures 6a and 6b.

[19] The largest discrepancy between the observations and LIMA in view of the influence of the solar activity occurs in the upper domain. In winter the influence of the solar activity upon the real water vapor values penetrates deeper by about 10 km than is reflected in the model calculations. However, the interruptions of measurements occurred in winter, making the statistics poorer. In summer the positive correlation also starts about 10 km deeper. We must note that the influence of the solar activity is separated from the general trend which additionally depends on different impacts such as change of the middle atmospheric dynamics, the alteration of exchange processes between troposphere and stratosphere, or the change of the concentration of

methane, etc. Nevertheless, the analysis is also influenced by uncertainties in both the calculations and observations. The vertical wind likewise changes with solar activity and is a sensitive parameter for the transport of water vapor. A somewhat too strong upward wind in summer could lift the domain of positive correlation upward. On the other hand, the decrease of the values of relative change for the observed data at the upper border displayed in Figure 6a may indicate an increasing uncertainty of the measurements with growing height as mentioned above. The analysis revealed a clear difference of behavior between summer and winter season. Despite some differences between observations and calculations, the order of change with the solar activity amounts to only a few percent below 80 km in summer and below 60–70 km in winter. According to the calculations the impact of the solar activity increases strongly toward the mesopause (not shown in Figure 6b).

[20] Figure 6 indicates an influence of the solar activity on the mesospheric water vapor distribution. But it is difficult to distinguish between the impact resulting from the variation of the Lyman- α radiation and that part caused by the internal dynamics which is not influenced by the solar activity. As reported by *Randel et al.* [2006] and *Scherer et al.* [2008] the Brewer-Dobson circulation in the tropics changed abruptly after 2001, impacting the water vapor distribution in the lower stratosphere. *Bittner et al.* [2000] and *Höppner and Bittner* [2007] found a slowdown of the planetary wave activity also in middle latitudes. Precisely such a sudden decrease in the water vapor mixing ratio was also observed at ALOMAR determining the trend in the whole period. This sudden change in the middle atmospheric dynamics during and after the solar maximum influences the correlation analysis between water vapor and solar Lyman- α radiation. Above 80 km the dependence is negative for all seasons including the summer months. Deducing from HALOE, *Chandra et al.* [1997] found a variation of the water vapor concentration over a solar cycle by 30–40% at 80 km and only 1–2% in the lower mesosphere (60–65 km). Also from HALOE, *Hervig and Siskind* [2006] derived a clear anticorrelation between Lyman- α radiation and water vapor mixing ratio at high latitude at 80 km in summer, as was likewise derived from the ALOMAR observations.

5. Summary and Conclusions

[21] Water vapor observations were carried out in the mesosphere (40–80 km) at ALOMAR (69.29°N, 16.03°E), Norway, from the end of 1995 until the end of 2006 covering nearly one solar cycle. The monitoring started shortly before the minimum of the solar activity and ended shortly before the next solar minimum so that any bias due to unsymmetrical solar activity was widely reduced. The measurements show a decrease of the mixing ratios at all heights during this period. A similar decline was calculated by the GCM LIMA. Because LIMA uses lower boundaries for the water vapor mixing ratio at the hygropause not changing from year to year, the decreasing tendency must result from altered dynamics. Precisely this assertion has also been confirmed by *Randel et al.* [2006] and *Scherer et al.* [2008] who stated a change of the Brewer-Dobson circulation in the tropical stratosphere and a slowdown of the extratropical

planetary wave activity [Bittner *et al.*, 2000; Höppner and Bittner, 2007].

[22] The dependence of the water vapor mixing ratio on the Lyman- α radiation as the most important proxy for solar activity influencing the upper mesosphere/mesopause region is different between summer and winter. The Lyman- α radiation impacts the upper domain (upper mesosphere and above) mainly by dissociation of water vapor forming molecular hydrogen, whereas its influence decreases exponentially with decreasing height. The absorption signal of Lyman- α is transported downward by the downward directed vertical wind in winter. The behavior is reversed in summer due to the upward flow of humid air in summer. The analysis shows that water vapor is positively correlated with the Lyman- α radiation in the lower mesosphere/stratopause region in winter. This could result from the impact of SSWs which are more frequent and stronger under conditions of high solar activity independent of the phase of the QBO [Sonnemann and Grygalashvyly, 2007]. However, when sorting the SSWs according to the phase of the QBO, a significant dependence was discovered. At high solar activity (i.e., larger than the mean amplitude of 10.7 cm flux), SSWs occurred more often during the west wind phase and at low activity during the east wind phase [Labitzke, 1987]. In summer a positive correlation exists at least up to the middle mesosphere. A possible reason is the autocatalytic water vapor production under the condition of high solar activity. But likewise a sudden change of the middle atmospheric dynamics can cause a so-called non-sense correlation. The LIMA calculations exhibit a somewhat different response. The positive correlation in summer reaches up to 75–80 km. As the lifetime of water vapor is extremely large at this altitude (from months up to infinity) a slightly enhanced vertical wind in the model can cause such an effect. The differences below 80 km are relatively small, with exception of the wintry values above about 60 to 70 km, so that the agreement between observations and calculations is relatively good.

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