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# Greenhouse gas effects on the solar cycle response of water vapour and noctilucent clouds

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## Abstract

The response of water vapour ($H_2O$) and noctilucent clouds (NLCs) to the solar cycle are studied using the Leibniz Institute for Middle Atmosphere (LIMA) model and the Mesospheric Ice Microphysics And tranSport (MIMAS) model. NLCs are sensitive to the solar cycle because their formation depends on background temperature and the $H_2O$ concentration. The solar cycle affects the $H_2O$ concentration in the upper mesosphere mainly in two ways: directly through the photolysis and, in time and place of NLCs formation, indirectly through temperature changes. We found that $H_2O$ concentration correlate positively with the temperature changes due to the solar cycle at altitudes above about 82 km, where NLCs form. The photolysis effect leads to an anti-correlation of $H_2O$ concentration and solar Lyman-α radiation, which gets even more pronounced at altitudes below ~83 km when NLCs are present. We studied the $H_2O$ response to Lyman-α variability for the period 1992 to 2018, including the two most recent solar cycles. The amplitude of Lyman-α variation decreased by about 40% in the period 2005 to 2018 compared to the preceding solar cycle, resulting in a lower $H_2O$ response in the late period. We investigated the effect of increasing greenhouse gases (GHGs) on the $H_2O$ response throughout the solar cycle by performing model runs with and without increases in carbon dioxide ($CO_2$) and methane ($CH_4$). The increase of methane and carbon dioxide amplify the

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response of water vapour to the solar variability. The solar cycle response is reduced in the late solar cycle due to a smaller amplitude of Lyman-α variability in the second period. Applying the geometry of satellite observations, we find a missing response when averaging over altitudes of 80 to 85 km, where $H_2O$ has a positive and a negative response (depending on altitude) which largely cancel out. One main finding is that during NLCs the solar cycle response of $H_2O$ strongly depends on altitude. A negative correlation between $H_2O$ and Lyman-α is found in the NLC sublimation zone below an altitude of about 83 km, but a positive response is present at the altitudes above 83 km where NLCs form.

## 1. Introduction

The 11-year solar cycle significantly influences the upper atmosphere's temperature and water vapour ($H_2O$) concentration. $H_2O$ is one of the essential minor constituents in the mesosphere as it is the primary source of chemically active hydrogen radicals, influencing the chemistry of all other chemically active minor constituents (Brasseur and Solomon, 2005, Hartogh et al., 2010). $H_2O$ concentration plays an essential role in the noctilucent cloud's (NLC) formation. NLCs are located at about 83 km altitude, consist of water ice particles, and owe their existence to the cold summer mesopause region (~130K) at mid and high latitudes. NLCs, also called polar mesospheric clouds, are formed in an environment where small changes in background $H_2O$ and temperature can lead to significant changes in NLC properties (e.g., Thomas, 1996; DeLand et al., 2006; Shettle et al., 2009, Lübken et al., 2009).

In comparison to the lower atmosphere, little is known about the upper mesosphere/lower thermosphere (MLT, 75-110 km) due to a lack of observations at these altitudes. NLCs have been proposed as indicators of trends (Thomas & Olivero, 2001). Studying NLC properties provide insight in phenomena occurring at the altitude of NLC. The 11-year solar cycle has been considered to cause quasi decadal oscillation observed in NLCs (DeLand et al., 2003).

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NLCs are predicted to decrease during solar maximum due to increased heating and photolysis of $H_2O$ (Garcia, 1989). However, some recent studies strongly suggest that the response of NLCs to the solar cycle has been absent from 2002 to the present (Fiedler et al., 2011; DeLand & Thomas, 2015; Hervig et al., 2016; Siskind et al., 2013). Hervig et al. (2019), using satellite observations, found that NLC had a clear anti-correlation with the solar cycle before 2002, and that response has been absent in recent years. The leading cause of this absence appears to be the suppression of the solar cycle response of $H_2O$. Lyman-$\alpha$ (Ly$\alpha$) radiation is the primary cause of $H_2O$ photolysis and varies by a factor of two between solar minimum and maximum (Woods et al., 2000). Understanding the effects of the solar cycle on $H_2O$ is more complicated at NLC altitudes because of the interaction between NLCs and background $H_2O$.

NLC growth leads to dehydration at higher altitudes (83-89 km) as ice particles are formed by consuming background $H_2O$, and sublimation of ice particles leads to hydration at lower altitudes as $H_2O$ is released here (about 78-83 km) (Lübken et al., 2009, Hervig et al., 2003). Investigating the effects of NLC on the background $H_2O$ requires an estimate of the $H_2O$ profile without NLCs. Investigations using satellite observations are limited due to uncertainty in the inferred background $H_2O$ without NLC and vertical resolutions in the order of a few 100~m. Therefore, using satellite observations to study $H_2O$ at NLC altitudes could yield misleading results due to biases in the estimated $H_2O$ profiles without NLC (Hervig et al., 2015). Hervig et al. (2015) suggest that in future studies, one approach to investigate the effects of NLC on $H_2O$ would be to use a detailed microphysical NLC model. NLC model simulations are performed with and without microphysics using the same background conditions, resulting in a $H_2O$ profile without NLC. This allows us to investigate how NLC formation changes the $H_2O$ background profile in detail.

We compare the model result to satellite observations published by Hervig et al. (2019) to investigate the mechanism behind the solar cycle response of NLC and $H_2O$. We also focus on the missing solar cycle response of $H_2O$ during recent years. This paper aims to answer a number of questions: How does the formation of NLCs affect the $H_2O$ profile and the variation of water vapour with the solar cycle? How do the solar cycle-induced temperature and photolysis changes affect the $H_2O$ response? Why is the response of water vapour to solar cycle nearly absent in satellite observations after 2005 (Hervig et al., 2019)? Our study is focused on the core NLC period, i.e., July at 68±5°N. The following section describes the modelling framework of this study and discusses the various model simulations performed. The third section discusses the mechanisms behind the solar cycle $H_2O$ response, such as the separation of the solar cycle-induced temperature and photolysis effects on $H_2O$. Sections four and five explore the possible reasons behind the missing solar cycle response. Concluding remarks and a summary are given in the last section.

## 2. Model description and numerical experiments

### 2.1. Model

The modelling framework used in this study consists mainly of two components: the Leibniz Institute Middle Atmosphere (LIMA) model and the Mesospheric Ice Microphysics And tranSport (MIMAS) model (see Fig. 1). LIMA is a non-linear, global, 3D Eulerian grid-point model reaching from the troposphere to the lower thermosphere, which calculates winds and temperature and is well described in a number of papers (Berger, 2008; Lübken et al., 2013). The LIMA model in this study is nudged to reanalysis data NOAA-CIRES (National Oceanic and Atmospheric Administration-Cooperative Institute for Research in Environmental Sciences 20CR; Compo et al., 2011) in the lower atmosphere. The resulting winds and temperatures in the mesosphere and lower thermosphere (MLT) are then used in MIMAS. The MIMAS model run was performed for all years with background wind conditions and gravity wave forcing from a representative year (1976).

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MIMAS is a 3D Lagrangian transport model specifically designed for modelling ice particles in MLT region (Berger and Lübken, 2015). MIMAS calculate NLC parameters from 10 May to 31 August, and it is constrained from mid-latitudes to high latitudes (37°-90°N) with a horizontal grid resolution of 1° in latitude and 3° in longitude and a vertical resolution of 100 m from 77.8 to 94.1 km (163 levels). In this study, the dynamics calculated by LIMA, solar Lyα, and the initial $H_2O$ distribution are the input for MIMAS as sketched in Figure 1. Two effects determine the mixing ratio of $H_2O$ in the mesosphere: (i) transport of $H_2O$ from lower altitudes and (ii) oxidation of methane ($CH_4$). The oxidation of each $CH_4$ molecule produces two $H_2O$ molecules. Methane is nearly completely converted to $H_2O$ in the mesosphere by photochemical processes (e.g., Lübken et al., 2018). MIMAS assumes that the $H_2O$ mixing ratio transported from lower altitudes is constant. Therefore, $H_2O$ concentration varies only according to $CH_4$ concentration (Lübken et al., 2018). Mesospheric $H_2O$ in MIMAS is transported by background winds, dispersed by turbulent diffusion, and reduced by photolysis. MIMAS makes use of 40 million dust particles, which can act as condensation nuclei. These are then coated with ice in $H_2O$ supersaturated regions and transported according to three-dimensional and time-dependent background winds, eddy diffusion, and sedimentation. In MIMAS, standard microphysical processes such as the Kelvin effect determine the nucleation and growth of ice particles (Berger & Lübken, 2015; Gadsden & Schröder, 1989). For the comparison with satellites, we used model run A, which includes $CO_2$ and $CH_4$ variations (Lübken et al., 2018; Lübken et al., 2021). We performed MIMAS model simulations with ice formation turned off and on respectively to investigate the effects of ice formation on background $H_2O$. In both runs, the background conditions and model inputs are the same. The main outputs of the model are the microphysical properties of the NLC ice particles, such as radius, backscatter value, and the number density of the ice/dust particles. More detailed descriptions of the MIMAS model and its precursors are available in the literature (Berger and

Fig. 1

von Zahn, 2002; Berger, 2008; Berger and Lübken, 2011; Lübken et al., 2018; Lübken et al., 2021).

## 2.2. Model simulations

LIMA and MIMAS use daily Lyα fluxes as a proxy for solar activity from 1961 to 2019. Lyα (and other bands) variations in LIMA cause atmospheric temperature variations, while Lyα variations in MIMAS cause photolysis of $H_2O$. In LIMA, variation of other bands, namely, Chappius band, Huggins band, Hartley band, Schumann-Runge band, and both Schumann-Runge continiums are parametrised following the Hamburg Model of the Neutral and Ionized Atmosphere (Schmidt et al., 2006) according to Lean et al. (1997). Therefore, it is possible to study the effects of solar cycle on $H_2O$ due to temperature changes and photolysis separately by performing model simulations with constant and varying Lyα in MIMAS and LIMA. We conducted four model runs, as described in Table 1. We also performed LIMA model simulations with constant $CO_2$ for runs E, F, and G to filter out their effects on temperature changes. For these runs we use a constant $CH_4$ concentration in MIMAS to avoid its influence on the $H_2O$ profile.

In LIMA, the mixing ratios of $CO_2$ (28-150 km) vary as function of time (years), while all other trace gases are kept constant. An increase in $CO_2$ leads to a decrease in temperature in the stratosphere mainly due to enhanced cooling by $CO_2$ (e.g., Roble and Dickinson, 1989; Garcia et al., 2007; Berger & Lübken, 2011; Marsh et al., 2013; Lübken et al., 2013). At NLC altitudes, this cooling leads to an altitude decrease of pressure levels, referred to as the ‘shrinking effect’ (Lübken et al., 2009). For LIMA we use the long-term increase of $CO_2$ concentration according to observations at Mauna Loa (19ºN, 155ºW).

This study focuses mainly on the recent two solar cycles from 1992 to 2018. Figure 2 shows the time series of Lyα, $CO_2$, and $CH_4$ for 1992-2018. The corresponding values of Lyα, $CH_4$, Fig. 2

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and $CO_2$ for the years considered for this study are highlighted. We classify 1992-2005 as period 1 ("early") and 2005-2018 as period 2 ("late"). Satellite observations of $H_2O$ showed a clear anti-correlation with the solar cycle in the early period, which was absent in the late period (Hervig et al. 2019). To investigate this, we first examined the early period solar minimum (1997) and maximum (2002) in more detail. The solar cycle affects the $H_2O$ concentration in two main ways. (i) through the photolysis of $H_2O$ by Lyα, and (ii) through the temperature effect. We distinguish these effects by performing model simulations with different background conditions (see Table 1). Namely in section 3.3, we discuss the individual role of solar cycle-induced photolysis and temperature change on the $H_2O$-solar cycle response. Figure 2 shows that the intensity of Lyα radiation during the late period has decreased compared to the early period, and the concentrations of increased greenhouse gases (GHGs) have increased in the late period. The effects of reduced Lyα intensity and increased greenhouse gas (GHG) concentration on long-term $H_2O$-solar cycle response are discussed in section 4.

## 3. Results and Discussions

### 3.1. Solar cycle response in ice water content (IWC)

To determine if the model agrees with satellite observations, we compared the ice water content (IWC) anomaly from the model with the satellite observations (see Fig. 3). IWC anomalies are calculated as follows: Fig. 3

$$IWC_{anom} = 100\% \cdot \frac{\overline{IWC_{July}} - \overline{IWC_{1981-2018}}}{\overline{IWC_{1981-2018}}}, \quad (1)$$

Where $\overline{IWC_{July}}$ represent monthly zonal averages at 68°N, and $\overline{IWC_{1981-2018}}$ are the averages of $\overline{IWC_{July}}$ over the years 1981-2018. The IWC anomaly for satellite measurements are from the Solar Backscatter Ultraviolet (SBUV), Halogen Occultation Experiment (HALOE), Cloud

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Imaging and Particle Size (CIPS), and Solar Occultation For Ice Experiment (SOFIE) instruments. The time series of SBUV and HALOE data as shown in Figure 3 represent three years sliding averaged values. For more details on the satellite datasets, see Hervig et al. (2019). For this comparison, we used the MIMAS run A, in which the simulations are performed with increasing concentrations of $CO_2$ and $CH_4$. For the comparison, we applied the same calculation method to our model data as Hervig et al. (2019) did on satellite observations, namely, we used a threshold of 50 g/km$^3$ for integrated water content because the PMC detection threshold for SBUV is 50 g/km$^3$ (DeLand and Thomas, 2015, 2019).

We find an anti-correlation between MIMAS IWC anomaly and Lyα flux throughout the entire period (1981-2018), with a weaker response in the late period. In satellite observations, SBUV measurements also show an anti-correlation with Lyα flux until 2005, after which the response becomes weaker in agreement with MIMAS. The magnitude of the solar cycle IWC anomaly in SBUV and HALOE is of the same order as the IWC anomaly in MIMAS. The IWC anomalies of CIPS and SOFIE do not show a clear response to the solar cycle. We notice that the year-to-year IWC variation in CIPS and SOFIE is larger than the IWC modulation during a solar cycle. IWC anomalies of SBUV and HALOE correlate well with MIMAS IWC anomalies before 2005 and progressively weaken afterwards. Lübken et al. (2009) found a good agreement between NLC parameters calculated by MIMAS and satellite observations. The general agreement between the main characteristics and trends of the ice layers in MIMAS and the observations suggests that the microphysical and photochemical processes in MIMAS cover the main processes relevant to NLC formation (Lübken et al., 2009).

### 3.2. Effect of NLC on water vapour ($H_2O$)

We calculated the zonal mean monthly averaged vertical profiles of $H_2O$ and temperature to investigate the impact of NLC formation on the $H_2O$ profile. Figure 4 shows the vertical $H_2O$ Fig. 4

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profile averaged for July at 68°N latitude and given on pressure altitudes $z_p = H_p ln(p_0/p)$. Where $p$ is the pressure of the model level, $p_0$ is the pressure at the surface and $H_p = 7$ km is the pressure scale height. This figure illustrates the effect of NLC formation on the background profile of water vapour since the $H_2O$ profile with NLC differs from that without NLC. In the presence of NLC there is a reduction in water vapour mixing ratio (dehydration) between 83-90 km, i.e. in the region where the saturation ratio of water vapour is larger than one. An enhancement in water vapour (hydration) is observed at altitudes between 79-83 km, where the saturation ratio of water vapour is smaller than one. An environment with a water vapour saturation ratio larger than one is supersaturated, meaning ice particles can grow under these conditions whereas a saturation ratio lower than one leads to ice sublimation. The degree of saturation depends on the background atmosphere's $H_2O$ concentration, and temperature. Ice particles formation starts at higher altitudes, where the temperature is the lowest, and then they sediment downward. During sedimentation, the ice particles grow by consuming $H_2O$ from the surrounding background, which decreases background $H_2O$ concentration. Then they approach a region with a saturation ratio smaller than one, where they sublimate, releasing the water vapour. This is the so-called freeze-drying effect well discussed in a number of papers (Hervig et al., 2003; Lübken et al., 2009; Bardeen et al., 2010). The results in Figure 4 also indicate that the effects of NLC on $H_2O$ are not present below ~79 km and above ~97 km. This is because the photochemical lifetime of water vapour below ~79 km becomes larger than dynamical characteristic times, and distributions of water vapour become dynamically determined. Above 97 km, the saturation ratio of water vapour is smaller than one; consequently, there is no NLC formation and consequently no effect on water vapour.

### 3.3. Effect of solar cycle-induced temperature and photolysis changes on water vapour ($H_2O$)

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We investigate the temperature change between the solar minimum (1997) and maximum (2002) due to solar irradiance variation and how these change affect the $H_2O$ profile. Different model runs performed for this study are summarized in Table 1. The differences (solar maximum - solar minimum) for $H_2O$ and temperature profiles are shown in Figure 5 for three model runs, namely E, F, and G. In run E, the solar cycle-induced temperature change and photolysis influences $H_2O$ concentration. In run F, only the temperature change caused by the solar cycle affects the $H_2O$ concentration, while in run G, only the photolysis caused by the solar cycle affects the $H_2O$ concentration (see Table 1). All of these runs are performed with constant $CO_2$ and $CH_4$ concentrations to avoid the effects of increasing GHG concentrations on temperature and $H_2O$ profiles.

In model run F, Lyα is held constant in MIMAS, so the photolysis of $H_2O$ is constant during the solar cycle. However, Lyα (and other bands) varies in the LIMA model, so the background temperature varies with the solar cycle. Therefore, the change in the $H_2O$ profile during the solar cycle is only due to the influence of the solar cycle on temperature and sequentially on microphysical processes. Figure 5a shows that the temperature increases during solar maximum compared to solar minimum through the entire altitude range (79-97 km). The difference of temperature amounts to ~0.5-1.7 K with maximum values at ~95 km. During solar maximum, increased solar irradiance leads to greater absorption of solar radiation in the MLT region by molecular oxygen and water vapour, which heats the background atmosphere. Temperature differences decrease as altitude decreases because the intensity of solar radiation decreases due to atmospheric absorption. The solar cycle effect in the $H_2O$ profile with NLC (blue line) differs significantly from that without NLC (yellow line). Without NLC, the $H_2O$ profile difference is nearly zero at all altitudes, indicating that the temperature changes do not significantly affect the background $H_2O$ profile in the absence of NLC. With NLC, the $H_2O$ profile difference is positive in the altitude range of 82-87 km and slightly negative in the range from 79 to 82 km.

Fig. 5a

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The atmosphere is warmer during solar maximum; therefore, the ice formation rate is lower during solar maximum. When the ice formation rate decreases, the amount of water vapour consumed from the background decreases; hence, more $H_2O$ is left in the background during solar maximum compared to solar minimum, resulting in a slightly positive response at NLC forming altitudes above 83 km. Below that altitude, the slightly negative response is due to reduced ice formation in the nucleation region during solar maximum, which decreases $H_2O$ released at ice sublimation altitudes. The positive difference peak at ~83 km is located near the bottom of the $H_2O$-saturated zone. Ice formation/sublimation is more sensitive to an increase in background temperature at this zone (where the degree of saturation is close to one) because at these altitudes the background temperature is almost equal to the frost point temperature, so an increase in background temperature critically changes the degree of saturation. The change of the background temperature in a region where it is significantly lower than the frost point temperature is not critical for the degree of saturation. Overall, the temperature variation due to the solar cycle causes a positive $H_2O$ response on the solar cycle at ice-formation altitudes and a slightly negative response at ice-sublimation altitudes.

In model run G (Fig. 5b), we consider only the effect of solar cycle-induced Lyα variation on water vapour photolysis. The background temperature is held constant. Photolysis of $H_2O$ by Lyα radiation molecules mainly leads to atomic hydrogen (H) and hydroxyl (OH) in the upper atmosphere (~90 %) and with less extent to $O(^1D)$ with molecular hydrogen (~10 %). The photolysis rate is higher during solar maximum due to the increased Lyα flux caused by the increased solar activity. Without NLC, the difference in the $H_2O$ profile is negative at all altitudes (yellow line), indicating that the background $H_2O$ is reduced during solar maximum due to increased photolysis. Figure 5b shows that the negative response peaks at an altitude of ~87.5 km. The solar cycle effect on the photolysis of $H_2O$ decreases above 87.5 km because the water vapour mixing ratio decreases with increasing altitude. The solar cycle variation of the photolysis effect decreases below 87.5 km because the solar Lyα radiation intensity decreases.

Fig. 5b

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With NLC (blue line), the $H_2O$ difference between the solar maximum and the solar minimum is essentially negative at ice sublimation altitudes (below ~83 km) and negligible at higher altitudes (above ~85 km). This is due to the redistribution of the $H_2O$ profile during NLC formation ("freeze-drying"). During solar maximum, the background $H_2O$ concentration available for ice formation is reduced due to enhanced photolysis. The lower $H_2O$ availability during solar maximum results in lower ice formation and, thus, lower $H_2O$ release during sublimation, leading to lower hydration in the sublimation zone. For this reason, the solar cycle variation of the photolysis effect is more pronounced at sublimation altitudes. Above 85 km, the effect of photolysis, in the case with NLC, is minimal because of the lower availability of $H_2O$ due to dehydration by NLC.

Figure 5c shows a combination of both effects, namely the solar cycle-induced temperature change and photolysis effects on $H_2O$. Without NLC (yellow line), the $H_2O$ profile shows a negative response at all altitudes, peaking at ~87.5 km similar to run G (Fig. 5b, yellow line). We found that the variation of temperature has an almost negligible effect on the $H_2O$ in the absence of NLC (see Fig. 5a, yellow line), so the negative response of water vapour without consideration of microphysical processes (yellow line on Fig. 5c) is mainly caused by the photolysis effect. With NLC (Fig. 5c, blue line), the combined effect of temperature and photolysis has a slightly positive response on water vapour in the ice formation zone (83-89km) and a negative response in the ice sublimation zone (80-83km). The slightly positive response is caused by the temperature modulation, and the negative response is primarily due to the photolysis modulation throughout the solar cycle. Fig. 5c

The study proves that the water vapour response to the solar cycle is affected by the re-distribution of water in the presence of NLC. There may exist regions with positive correlation of water vapour with Lyα when NLC formation occurs. Without NLC, the water vapour always shows a negative correlation to the solar cycle. When comparing the effects of solar cycle modulations of temperatures and photolysis on $H_2O$, the photolysis has a stronger effect on

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water vapour, however, the variation of temperature induces a positive correlation of solar irradiance and $H_2O$.

## 4. Increasing greenhouse gases and reducing solar cycle

This section examines how the increase in GHGs affects the $H_2O$ response to the solar cycle. To distinguish the GHG effects, we compared the model results with increasing $CO_2$ and $CH_4$ (Run A) to the model run with constant $CO_2$ and $CH_4$ (Run E). It is noted already that an increasing $CO_2$ concentration leads to a cooling of the middle atmosphere, and an increase in $CH_4$ concentration leads to an increase in $H_2O$ concentration (see Sec.2 for details). In Figure 2, the concentration of $CO_2$ and $CH_4$ increase during the late period, and at the same time, the peak of the Lyα flux decreases. In order to filter out the effect of reduced Lyα intensity, we calculated the $H_2O$ response profile per unit of Lyα ($\Delta H_2O / \Delta Ly\alpha$). Figure 6 shows the result for the first (1997-2002, blue line) and the second period (2008-2014, orange line) for model runs E (Fig. 6a) and A (Fig. 6b) respectively. These profiles show positive and negative responses depending on altitude. Under the conditions of constant GHGs (run E) the sensitivity of water vapour to Lyα does not change from the early to the late period (Fig. 6a). As expected, for the case of growing methane and carbon dioxide (run A), the sensitivity of water vapour to Lyα increases during the late period (orange line, Fig. 6b) compared to the early period (blue line, Fig. 6b). This is because an increase in $CO_2$ (and consequently temperature decrease) leads to an intensification of microphysical processes, hence, to the increased freeze-drying. In addition, increasing methane leads to more water vapour in the upper mesosphere, which also leads to an increased water vapour variation with solar cycle. Fig. 6

To study the effect of a decreasing Lyα amplitude during the late period (2008-2014), we calculated the ratio of water vapour absolute deviations between solar minimum and solar

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maximum for the early and late period. The amplitude of Lyα variation is weaker during the late period (~$1.14 \cdot 10^{11}$ [phot.cm$^{-2}$s$^{-1}$]/solar cycle) compared to the early period (~$1.85 \cdot 10^{11}$ [phot.cm$^{-2}$s$^{-1}$]/solar cycle). The intensity of Lyα during the late period solar maximum is reduced by ~40% compared to the early period. As can be seen from Figure 7a the magnitudes of positive and negative $H_2O$ responses decreased during the late period for model runs with constant GHGs (Run E). In Figure 6a, we found that the $H_2O$ sensitivity to Lyα flux is the same in the early and late periods for the model run with constant GHGs (Run E). Therefore, the reduced response of $H_2O$ during the late period in model run E (Fig. 7a) is only due to the reduced solar Lyα variation. Comparing the late period $H_2O$ response to the solar cycle from model runs with constant GHG (Fig. 7a, orange line) to model runs with increasing GHG (Fig. 7b, orange line) suggests that both the positive and negative peak responses are enhanced by increasing GHG concentration.

Fig. 7

## 5. Missing $H_2O$-solar cycle response

A recent study by Hervig et al. (2019) reported a missing response in $H_2O$ concentration on solar cycle after 2005. In Figure 8, we compare our model results of $H_2O$ anomaly with the satellite observations. The $H_2O$ response is averaged over the geometric altitudes of 80-85 km at 68°N. For this comparison, we used MIMAS run A, where the increasing concentration of GHG is considered. The satellite observations are shown in Figure 8 from HALOE, SOFIE, and MLS according to Hervig et al. (2019). HALOE shows a strong negative response to Lyα (-1.7 ppmv/solar cycle) during period 1, but in SOFIE and MLS the response is almost absent (+0.2 ppmv/solar cycle) during period 2 (Hervig et al., 2019). For MIMAS, no clear $H_2O$-solar cycle anti-correlation is noticed in the early period, but it was slightly positive in the late period in agreement with SOFIE and MLS satellite observations. To investigate the $H_2O$ response on

Fig. 8

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Lyα variation in more detail, we analysed the vertical $H_2O$ response profile at geometric altitudes similar to the satellite observations.

Figure 9 shows the vertical profile of $H_2O$ response in geometric altitudes for the model run with constant GHGs (run E, Fig. 9a) and growing GHGs (run A, Fig. 9b). The magnitude of the $H_2O$ response at geometric altitudes (Fig. 9) differs from that at pressure altitudes (Fig.7). This is because the geometric altitude of constant pressure levels is not constant and varies throughout the solar cycle but also with time due to increasing GHG. Therefore, the magnitude of the $H_2O$ response differs when converted from pressure altitudes to geometric altitudes. Fig. 9

We focus on the 80-85 km geometric altitude range (Fig. 9. shaded region). There are positive and negative $H_2O$ response zones within this altitude range similar to Figure 7. We calculated the average $H_2O$ response over 80-85 km altitude range for MIMAS runs A and E is given in Table 2. For the model run with growing GHGs (run A), the $H_2O$ response averaged over an altitude range of 80-85 km changed from -0.01 ppm/solar cycle in the early period to 0.10 ppm/solar cycle in the late period (see Table. 2). The $H_2O$ response in the late period becomes slightly positive for run A, consistent with the satellite observations of SOFIE and MLS (see Fig. 8). The vertical profile of $H_2O$-solarcycle response clearly show that H2O response to the solar cycle is not completely missing in the late period. The missing response in the MIMAS $H_2O$ as shown in Figure 8 occurred when averaging over the 80-85 km altitude range. Figure 9 demonstrates that he $H_2O$ response shows nearly equal positive and negative responses within the 80-85 km altitude range (shaded region). Therefore, averaging the response in this altitude range becomes nearly zero, as the positive and negative responses cancel out each other. When averaging over the altitude range of 80 to 82 km in the early period we receive $H_2O$ response of -0.71ppm/solar cycle and an anti-correlation between $H_2O$ and Lyα . The results clearly shows that the small solar cycle response in MIMAS is a consequence of averaging over an altitude range of 80-85 km. It suggests that averaging $H_2O$ response over an altitude range

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containing positive and negative responses may not provide a detailed understanding of the $H_2O$-solar cycle response.

## 6. Conclusions

In this study, we used our ice particle model MIMAS along with atmospheric transport model LIMA to investigate the response of $H_2O$ to the solar cycle from 1992-2018. We investigated how NLC formation affects vertical $H_2O$ profiles by running model simulations with and without microphysics. NLC formations are shown to redistribute $H_2O$ profiles by consuming $H_2O$ from the background at ice-forming altitudes (dehydration) and releasing it at ice-sublimating altitudes (hydration) is known as the "freeze-drying" effect. To investigate the missing solar cycle response in satellite observations reported by Hervig et al. (2019), we divided the entire study period into an early (1992-2005) and late (2005-2018) period. We first investigated how the Lyα variation affects the $H_2O$ profile between solar minimum and maximum in the early period. The solar Lyα variation affects the $H_2O$ concentration at NLC altitudes mainly in two ways: through the effect of temperature change and through the effect of photolysis. To distinguish these two effects, we performed additional model simulations with different background conditions (see Table 1). We found that the modulation of water vapour, which comes through the temperature changes with solar cycle, causes a slight positive $H_2O$ response at ice-forming altitudes and a negative response at ice-sublimating altitudes. The solar cycle photolysis effect has only negative responses on the $H_2O$ profile, and this response dominates at ice sublimation altitudes with NLC conditions. Our results for the case of photolysis effect only are supported by previous simulations, which also suggest that freeze-drying significantly reduces the potential effect of Lyα photolysis on $H_2O$ above 82 km, while the effect is enhanced at 80-82 km, where ice particles sublimate (von Zahn et al., 2004, Lübken et al., 2009).

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To the best of our knowledge, we have for the first time identified a positive response of water vapour to Lyα variation in the MLT region which is due to microphysical processes. It was assumed for a long time that water vapour only anti-correlates with the solar cycle at mesopause altitudes (e.g. Sonnemann and Grygalashvyly, 2005; and references therein). We should note that in the Martian atmosphere, where microphysical processes have a crucial role in water vapour distributions through the entire atmosphere in all seasons (e.g. Shaposhnikov et al., 2018), this effect may be important.

We have made a comparison between the model and satellite observations of the $H_2O$ response to the solar cycle, averaged over an altitude range of 80-85 km. The satellite observations from HALOE show a strong anti-correlation to the solar cycle in the early period, but the model shows a very small response in both the early and late periods. The vertical $H_2O$ response profiles from MIMAS show that within the 80-85 km altitude range, the positive and negative responses are almost equal in magnitude and symmetric. Therefore, averaging the response over this altitude range reduces the overall response in model, as positive and negative responses cancel each other out.

We also investigated the role of increasing GHG on the $H_2O$-solar cycle response. From the early to the late period, there are mainly two factors that affect the long-term $H_2O$ solar cycle response: increasing $CO_2$ and $CH_4$ concentrations and the lower intensity of the solar cycle (see Figure 2). We found that increasing GHG concentration increased the $H_2O$ response to Lyα. The Lyα intensity during the late solar maximum decreased by 40% compared to the early solar maximum. Therefore, the overall response of $H_2O$ to the solar cycle is also decreased in the late period.

**Appendix**

**Data availability.** The data utilized in this manuscript can be downloaded from

**Author contributions.** All authors contributed equally to this paper.

**Competing interests.**

**Acknowledgements.** We acknowledge the Mauna Loa records for $CO_2$ and $CH_4$ from http://www.esrl.noaa.gov/gmd/ccgg/. This paper is partly supported by the TIMA project of the BMBF research initiative ROMIC.

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# Figures

Figure 1. Sketch of the LIMA (green) and MIMAS (blue) models (from Lübken et al., 2021)

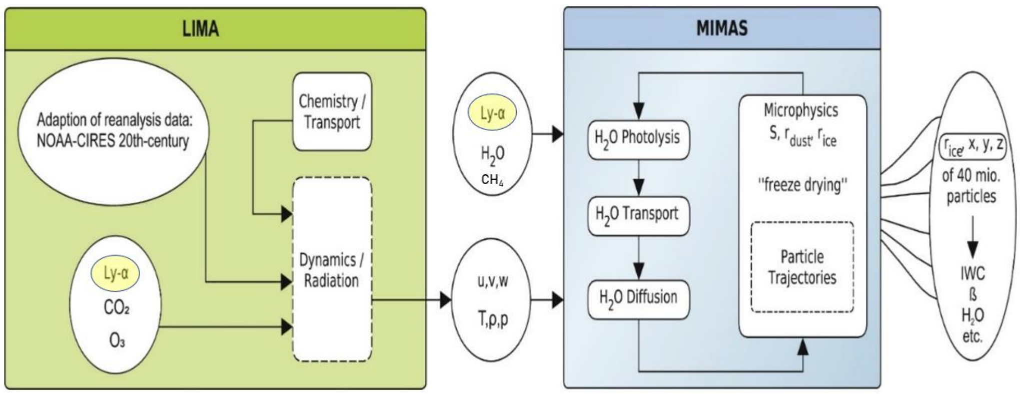

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Figure 2. Time series of solar Lyα, $CO_2$, and $CH_4$ for 1992-2018. The corresponding Lyα, $CO_2$, and $CH_4$ values for the solar cycle maximum and minimum years used for this study are marked. The $CO_2$ and $CH_4$ values for run A are represented with dots, and for run E with crosses. The study period is divided into period 1 as early (1992-2005) and period 2 as late (2005-2018).

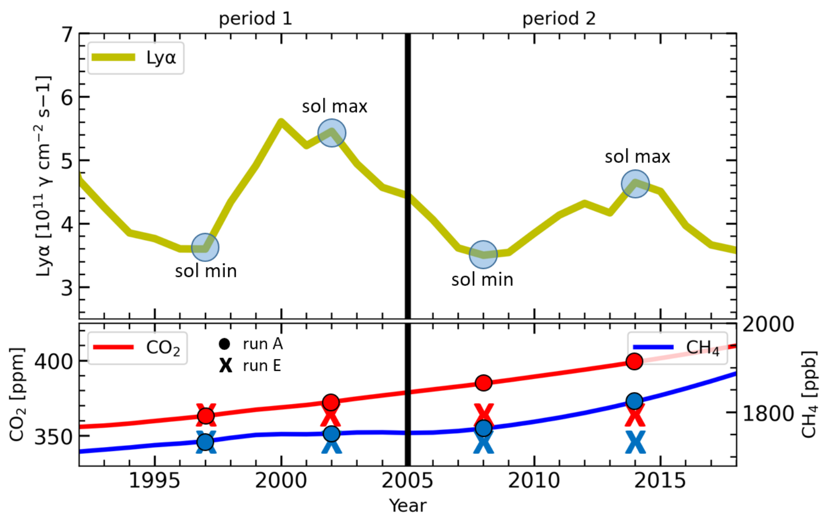

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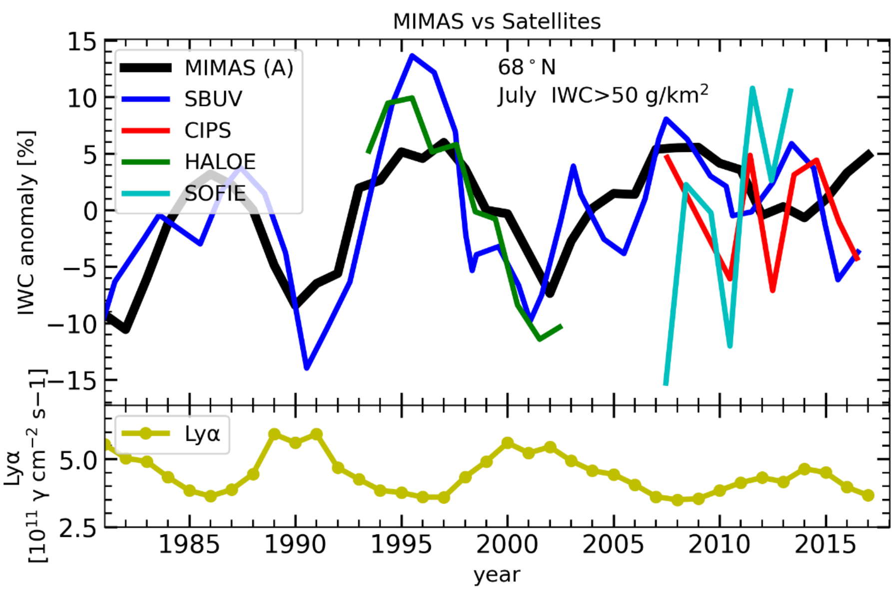
Figure 3. Time series of July mean IWC anomalies at 68°N from model and satellites based on Hervig et al., (2019). Anomalies for each data set are calculated as the difference from their long-term mean. To reduce year to year variability, the time series of SBUV and HALOE are smoothed using the sliding average method of window size 3. Lyα -solar cycle modulation is shown in the bottom panel.



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619 Figure 4. Zonally and monthly averaged $H_2O$ and temperature profile for July at 68°N from
620 MIMAS with and without NLCs. The red dotted line represents frost point temperature. The
621 blue lines show the background $H_2O$ concentration with NLC, and the yellow lines show the
622 $H_2O$ concentration without NLC.

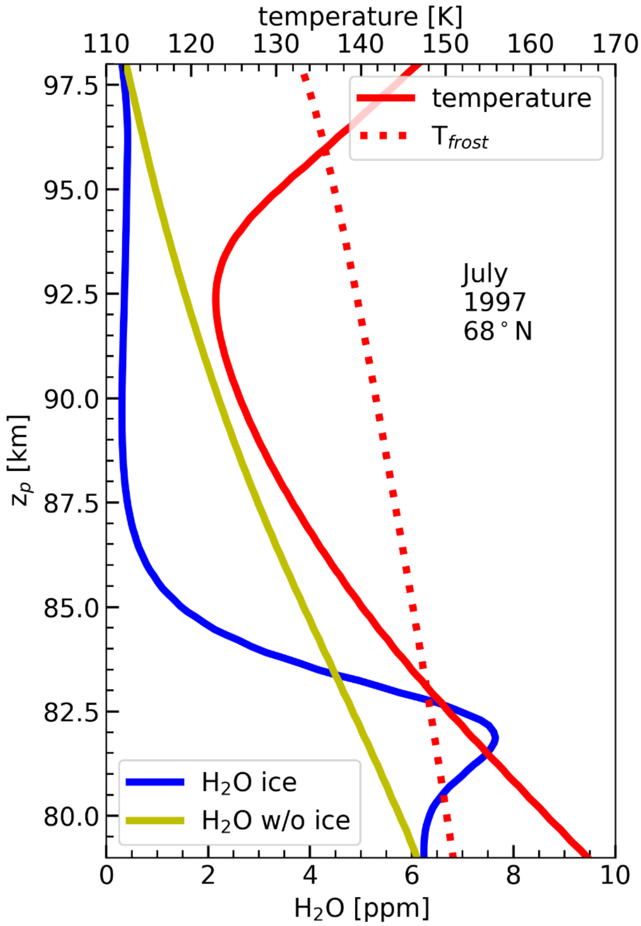


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Figure 5. The difference in profiles between solar maximum (2002) and minimum (1997) for July mean $H_2O$ and temperatures. The blue and yellow lines represent NLC and non-NLC conditions. In all cases, $CO_2$ and $CH_4$ values are constant corresponding to 1997. (a) Run F: only temperature change effects on $H_2O$, (b) Run G: only Photolysis change effect on $H_2O$, (c) Run E: both temperature change and photolysis change effects on $H_2O$

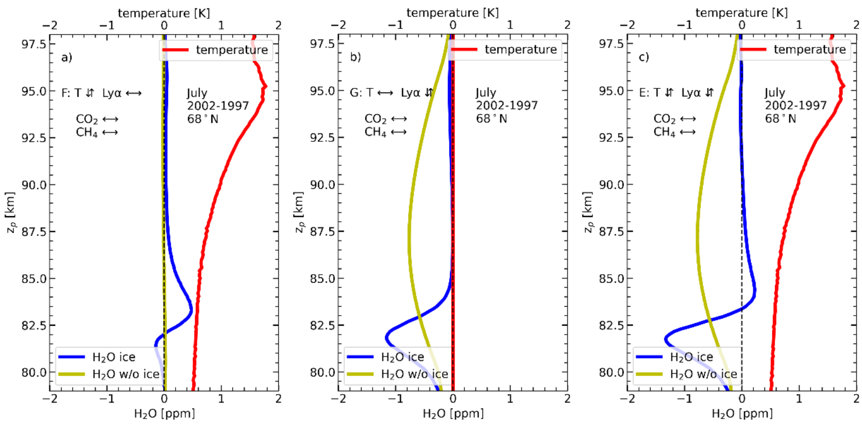

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Figure 6. $H_2O$-response per unit Lyα variations in July at 68°N during the years between solar minimum and maximum in the early (1997 to 2002) and late (2008 to 2014) periods. (a) MIMAS model run E with constant $CO_2$ and $CH_4$, (b) MIMAS model run A with varying $CO_2$ and $CH_4$.

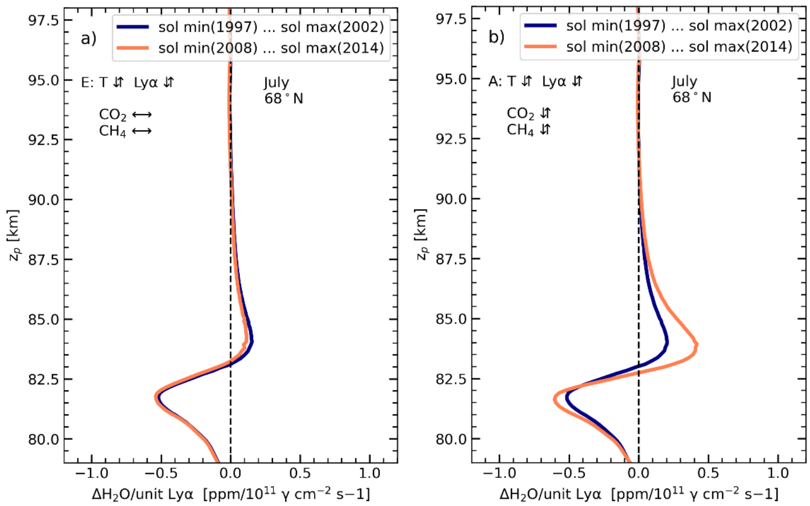

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Figure 7. $H_2O$-response to absolute solar cycle Lyα variations in July at 68°N during the years between solar minimum and maximum in the early (1997 to 2002) and late (2008 to 2014) periods. (a) MIMAS model run E with constant $CO_2$ and $CH_4$, (b) MIMAS model run A with varying $CO_2$ and $CH_4$.

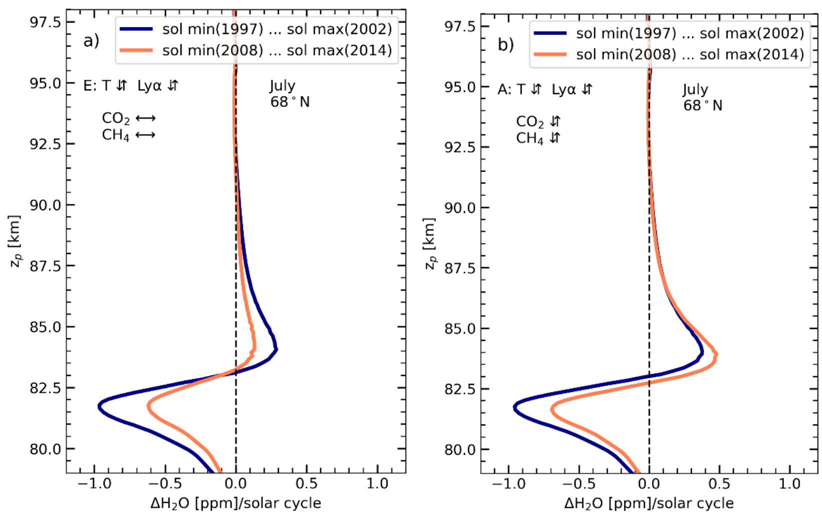

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Figure 8. Time series of Lyα and $H_2O$ anomalies as monthly averages for July at 68°N for the altitude range of 80 km to 85 km from MIMAS run A and satellites (HALOE and the composite data (MLS and SOFIE)). Satellite observations are according to Hervig et al., 2019. The $H_2O$-Lyα correlation is calculated for the early and late periods (see inlet).

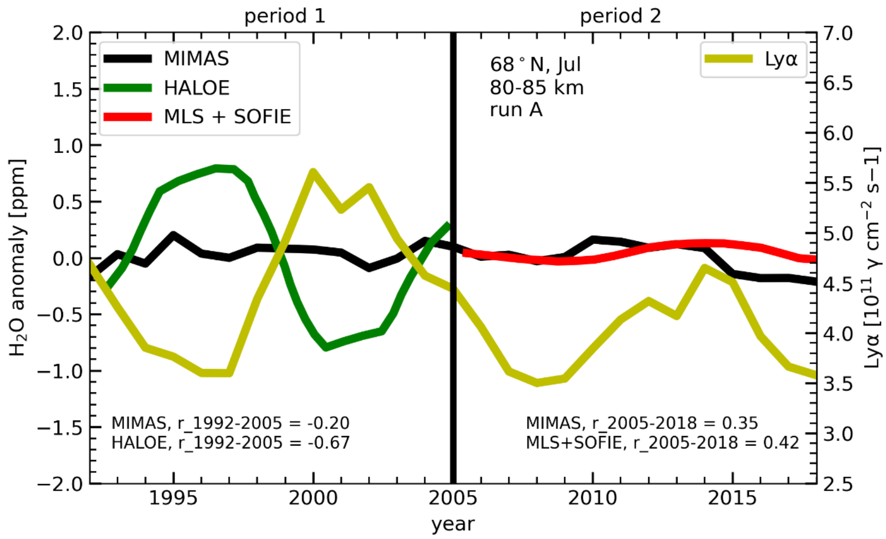
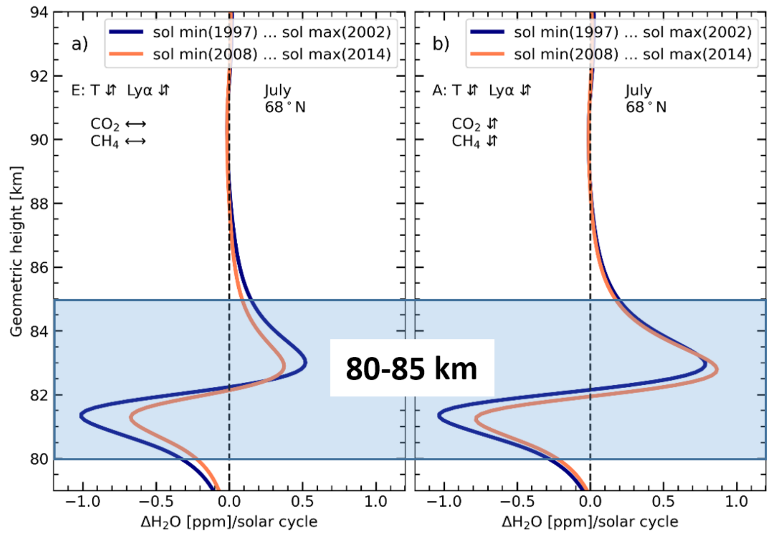


Figure 9. $H_2O$-response to absolute solar cycle Lyα variations in July at 68°N during the years between solar minimum and maximum in the early (1997 to 2002) and late (2008 to 2014) periods represented in geometric altitudes. The shaded region represents the altitudes range used for calculating an average solar cycle response. (a) MIMAS model run E with constant $CO_2$ and $CH_4$, (b) MIMAS model run A varying $CO_2$ and $CH_4$.



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## Tables

Table 1. MIMAS simulations were carried out under different background conditions. The horizontal arrow stands for constant values for the given year; the vertical arrow is for varying parameters. How Lyα affects $H_2O$ is given for each run in the last column.

| | **LIMA** | | **MIMAS** | | |
|---|---|---|---|---|---|
| **Model run** | **$CO_2$** | **Lyα T effect** | **$CH_4$** | **Lyα photolysis effect** | **Water vapour solar cycle response affected by** |
| **A** | ↕ | ↕ | ↕ | ↕ | • **Temperature change (Lyα + $CO_2$ )**<br>• **Photo dissociation**<br>• **Varying $CH_4$ ($H_2O$ source)** |
| **E** | ↔ **1997** | ↕ | ↔ **1997** | ↕ | • **Temperature change**<br>• **Photo dissociation** |
| **F** | ↔ **1997** | ↕ | ↔ **1997** | ↔ **1997** | • **Temperature change** |
| **G** | ↔ **1997** | ↔ **1997** | ↔ **1997** | ↕ | • **Photo dissociation** |

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Table 2. The solar cycle $H_2O$ response averaged over 80-85 km geometric altitude at 68°N for model runs A and E.

| **Model run** | **$\Delta H_2O$ (ppm)/solar cycle (80-85km)** | |
|---|---|---|
| | **Early period** | **Late period** |
| **MIMAS with constant $CO_2$ and $CH_4$ (run E)** | **-0.11** | **-0.06** |
| **MIMAS with increasing $CO_2$ and $CH_4$ (run A)** | **-0.01** | **0.10** |