Modeling of the middle atmosphere response to 27-day solar irradiance variability.

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Abstract

The solar rotational variability (27-day) signal in the Earth's middle atmosphere has been studied for several decades, as it was believed to help in the understanding of the Sun's influence on climate at longer timescales. However, all previous studies have found that this signal is very uncertain, likely due to the influence of the internal variability of the atmosphere. Here, we applied an ensemble modeling approach in order to decrease internal random variations in the modeled time series. Using a chemistry-climate model (CCM), SOCOLv3, we performed two 30-member 3-year long (2003-2005) ensemble runs: with and without a rotational component in input irradiance fluxes. We also performed similar simulations with a 1-D model, in order to demonstrate the system behavior in the absence of any dynamical feedbacks and internal perturbations. For the first time we show a clear connection between the solar rotation and the stratospheric tropical temperature time-series. We show tropical temperature and ozone signal phase lag patterns that are in agreement with those from a 1-D model. Pronounced correlation and signal phase lag patterns allow us to properly estimate ozone and temperature sensitivities to irradiance changes. While ozone sensitivity is found to be in agreement with recent sensitivities reported for the 11-year cycle, temperature sensitivity appears to be at the lowest boundary of previously reported values. Analysis of temperature reanalysis data, separate ensemble members, and modeling results without a rotational component reveals that the atmosphere can produce random internal variations with periods close to 27 days even without solar rotational forcing. These variations are likely related to tropospheric wave-forcing and complicate the extraction of the solar rotational signal from observational time-series of temperature and, to a lesser extent, of ozone. Possible ways of further improving solar rotational signal extraction are discussed.

Keywords

Middle atmosphere, 27-day solar cycle, temperature, chemistry-climate modeling

1. Introduction

The understanding of the physical/chemical mechanisms responsible for the influence of solar irradiance changes on climate is of significant interest and is currently an important scientific problem. Recent progress in this area is based on the analysis of new observations and model results, with the aim to better characterize the response of the entire atmosphere and climate system to solar irradiance variability on different time scales. The decadal-scale solar signal from a number of new meteorological reanalysis datasets has been presented by Mitchell et al. (2014). The authors confirmed previously discovered warming in both the upper and lower tropical stratosphere during solar activity maximum years. This double-peak structure is a part of the “top-down” mechanism of a solar-climate connection, currently an area of intensive research (see Gray et al., 2010 and references therein). Understanding of the mechanisms behind the warming in the upper stratosphere has been improved in the last decade and several state-of-the-art chemistry climate models are able to reproduce this feature (e.g., SPARC, 2010, chapter 8), although this is less robust in results from models participating in the IPCC CMIP-5 activity (Mitchell et al., 2015; Hood et al., 2015). The analysis of the CMIP-5 results also showed that the warming in the tropical lower stratosphere cannot be clearly identified in a majority of the models.

Understanding of the decadal-scale stratospheric response to solar irradiance variability is further complicated by problems with how it is extracted from the relatively short observational time series using multivariate linear regression analysis (MLRA). Chiodo et al. (2014) demonstrated that the lower tropical stratosphere warming extracted from the simulated temperature time series appears only in the model run driven by all forcing agents including stratospheric sulfate aerosols produced by volcanic eruptions. However, Chiodo et al. (2014) used a model with a prescribed ocean (WACCM), and there are reasons to think that feedbacks from the tropical troposphere/ocean response could be important in increasing the amplitude of the lower stratospheric response (e.g. Hood and Soukharev, 2012). Note that recent experiments with the upgraded WACCM including an interactive ocean already reveal a lower stratospheric warming due to the 11-year solar cycle obtained with MLRA during volcanically quiet periods (Lon Hood, November 2016, private communication). Further, the solar signal even in the tropical upper stratospheric temperature is shown to be very model dependent (Hood et al., 2015; Mitchell et al., 2015). Nevertheless, the conclusions of Chiodo et al. (2014) were partially confirmed by Mitchell et al. (2015), who showed that even in the models with coupled oceans the warming of the lower tropical stratosphere is much less pronounced for the longer period (1850–2005) compared to the period highly affected by volcanic activity (1979–2005). Austin et al. (2008) analyzed results of several chemistry-climate models for the period 1960–2005 and also showed that the ozone solar signal in the lower tropical stratosphere changes substantially from the first two solar cycles (1960–1981) to the last two (1982–2003), but attributed this issue to an aliasing between the sea surface temperatures and the solar cycle during the first part of the period. All this casts some doubts on the robustness of the lower stratospheric ozone and temperature responses to solar irradiance variability extracted from a relatively short time series using MLRA (Maycock et al., 2016).

One way to partially avoid the problem of robust statistics or long-term issues in observations, and

better understand the processes is to analyze the stratospheric response to solar irradiance variability on shorter time scales, such as intra-seasonal variability related to the solar rotational (27-day) cycle. Solar 27-day variability has similar spectral distribution of irradiance changes as the 11-year variability, since they are both related to the distribution of active regions on the Sun's surface (Yeo et al., 2015). The amplitude of 27-day cycles is less pronounced during the minimum phase of the 11-year cycle, but comparable to the amplitude of the 11-year cycle during the 11-year solar cycle maximum phase. Atmospheric responses to the 27-day solar variability have been studied for more than three decades using observations (e.g. Hood, 1984; Hood and Zhou, 1998; Ruzmaikin et al., 2007) and chemistry-climate 1-D (Eckman, 1986; Brasseur et al., 1987; Summers et al., 1990; Chen et al., 1997), 2-D (Fleming et al., 1995; Chen et al., 1997; Zhu et al., 2003), and 3-D (Williams et al., 2001; Rozanov et al., 2006; Gruzdev et al., 2009; Grenfell et al. 2010; Kubin et al., 2011) models (CCM).

In the mesosphere, a clear connection between the solar irradiance Lyman- α line and H₂O (Shapiro et al., 2012; Thomas et al., 2015; Lainer et al., 2016), OH (Shapiro et al., 2012), and HO₂ (Wang et al., 2015) has been found. A significant response to the rotational cycle was also shown for CO (Ruzmaikin et al., 2014) and for other species (Gruzdev et al., 2009). Analysis of the observed and modeled ozone (e.g., Bossay et al., 2015 and references therein) showed good agreement between observed and simulated stratospheric ozone variability. Typically, the ozone response consists of a well-pronounced positive correlation with increasing time lag: from slightly negative (several days) around the stratopause to positive (up to 10 days) in the lower stratosphere. The negative phase lag is present due to the ozone-temperature feedback, i.e. temperature dependence of the catalytic cycles of ozone destruction (e.g. Hood, 1986). In the mesosphere, the ozone response is in the opposite phase at a zero lag, which is due to a high importance of the HO_x (OH+HO₂) ozone-destroying cycle at these altitudes. The maximum stratospheric ozone response to irradiance changes is found around 40 km (0.2-0.6 % to 1 % change of irradiance at 205 nm (I205)). It was also shown that the ozone response strongly depends on the period considered (Hood and Cantrell, 1988; Hood and Zhou, 1998; Bossay et al., 2015), which suggests, first of all, that the forcing has to be high enough to cause observable changes, but also that the response can depend on the state of the system, since in some cases the less pronounced signal is shown for the periods which are closer to the solar max conditions and therefore have higher signal-to-noise ratio. Hood and Zhou (1998), however, also noted that such a discrepancy can be related to the instrumental issues in ozone observations.

A temperature response is much less robust. Although most of the reported maximum temperature sensitivities in the tropics (0.06 % or 0.16 K per 1 % change in I205, Hood, 1986; Rozanov et al., 2005; Gruzdev et al. 2009) were similar to those derived for the 11-year cycle case (1-1.5 K solar max/min difference, Frame and Gray, 2010; Mitchell et al., 2014), the correlation coefficients reported didn't exceed 0.2-0.3 in the middle atmosphere and, additionally, there was a high uncertainty in the reported phase lags. It was suggested (Chandra 1986, Rozanov et al., 2005; Gruzdev et al. 2009) that this disagreement might be related to the influence of atmospheric internal variability that makes extraction of the solar signal from the data more difficult. Using a 3-D CCM, Rozanov et al. (2006) provided evidence for this hypothesis by showing a high spread of results for different modeling ensemble members. Gruzdev et al. (2009) demonstrated a robust signal in modeled temperature, but only by

using a solar rotational forcing amplified by a factor of three. Recently, Garfinkel et al. (2015) applied superposed epoch analysis to newly available reanalysis data and demonstrated that the temperature response to 27-day solar irradiance variability significantly depends on the QBO phase. Since the QBO phase strongly modulates the wave propagation in the stratosphere, this finding highlights the important role of internal variability. Ruzmaikin et al. (2007) and Garfinkel et al. (2015) also found a strong, close to 27-day, periodicity in the winter polar stratosphere, which emphasizes that similar mechanisms can be responsible for the influence of solar irradiance on the atmosphere on both decadal and intra-seasonal time scales. On the other hand, in the tropics, where the temperature signal is expected to be the highest, it has not been well established, despite the use of many different statistical methods (Gruzdev et al., 2009 and references therein). Therefore, it is still unclear whether the reported polar oscillations are related to solar irradiance changes. The intra-seasonal variability of the polar vortex can still potentially be a result of internally generated oscillations, so-called stratospheric vacillation cycles or minor heating events (Holton and Mass, 1976), which tend to occur in the same months as reported by Ruzmaikin et al. (2007) and Garfinkel et al. (2015) and can be generated both by oscillations of tropospheric wave forcing and by the stratosphere itself (Scaife and James, 2001). Chandra (1986) showed that these dynamically induced oscillations have periods varying from 3 to 7 weeks and concluded that they appear to be independent of solar activity. Chandra (1986) also showed that the stratospheric low-latitude ozone and temperature is much more sensitive to dynamically induced changes than to changes in solar irradiance. Kuroda and Kodera (2001) characterized such short-term variability in the polar areas, using monthly mean data, as the Polar Night Jet Oscillation (PJO) and attributed its origin to the variability of tropospheric wave forcing, which is supported by the fact that the PJO is more frequent in the northern hemisphere. Kuroda and Kodera (2002), however, then also showed that the PJO can be a subject of modulation by the 11-year solar cycle.

In this study, we attempt to distinguish between the internal variability of the atmosphere in the tropics and the variability forced by the 27-day solar rotational cycle. We use 1-D modeling as an idealized example of how the atmosphere would be expected to respond without internal variability present, and 3-D ensemble modeling to elucidate the intra-seasonal solar related variability. Since the wave generating sub-synoptic dynamical activity in the troposphere is not perfectly represented by global models (e.g. Small et al., 2014), the related short-term oscillations in the model are random depending on initial conditions. A sufficient number of ensemble model runs should allow the role of random components to be minimized. We also analyze the latest observations and reanalysis data.

2. Description of datasets, models, experiments, and methods

2.1 SSI data

As a solar forcing we used the Naval Research Laboratory (NRL) SSI reconstruction (Lean, 2000). NRLSSI is an empirical model that uses the disk-integrated Mg II and photospheric sunspot indices to describe the evolution of sunspots and faculae, respectively. Above 400 nm, the reconstruction is based on the facular and sunspot contrasts calculated by Solanki and Unruh (1998) with the ATLAS9 code. The data below 400 nm are based on the UARS/SOLSTICE measurements and multiple regression analysis. The NRL SSI data was shown to be in a good agreement with other reconstructions and

observational data on the intra-seasonal scales, although there is a generally increasing uncertainty in solar variability above 300 nm (e.g. Yeo et al., 2015), which can be important for the atmosphere (Shapiro et al., 2011) but is not yet studied using 3-D models. For our analysis we used the years 2003-2005, the descending phase of solar cycle 23, which covers both near-maximum and near-minimum rotational variability, which allows us to study the evolution of the atmospheric response with respect to the strength of the forcing applied. As a proxy for cross-correlation analysis of atmospheric parameters we used the irradiance flux at 205 nm.

(Figure 1)

2.2 RCPM 1-D model

To illustrate the middle atmosphere response to 27-day SSI variability in the absence of any internally generated variability, we use a 1-D radiative-convective-photochemistry model (RCPM). The model was developed by Egorova et al. (1997) and Rozanov et al. (2002). It consists of radiation, chemistry, convective adjustment, and vertical diffusion modules and has 40 layers from 0 km to 100 km. The photochemical part of the model calculates the distribution of 43 chemical species of the oxygen, nitrogen, hydrogen, carbon, chlorine and bromine groups. To reach the equilibrium state we first ran the model for 50 years with a 2-hour time step using a constant irradiance flux with irradiance as at 1 January 2003. Then we ran the model for 3 years using the reference daily solar forcing illustrated in Figure 1. Cross-correlation functions of the model output time-series of temperature, ozone, and hydroxyl with respect to the irradiance proxy are analyzed in Section 3.1.

2.3 SOCOL 3-D model

As our main modeling tool we used the chemistry-climate model (CCM) SOCOL v3.0. It consists of the dynamical core MA-ECHAM5 (Roeckner et al., 2003) and the chemical core MEZON (Egorova et al., 2003), interacting with each other every 2 model hours. For the present study, we applied T42 horizontal resolution ($\sim 2.8^\circ \times 2.8^\circ$) and a vertical structure of 39 levels between Earth's surface and 0.01 hPa (~ 80 km). Dynamical processes in the CCM SOCOL are calculated with a 15 min time step, while full radiative and chemical calculations are performed every 2 hours. The detailed description of the model is given by Stenke et al. (2013), who also showed a reasonable representation of the climate state of the middle atmosphere compared to observations. The applied model version underwent some further improvements described by Revell et al. (2015). The main changes related to the representation of the solar effects are the application of extra-heating parameterization allowing a better simulation of the heating rate response to SSI variability (Sukhodolov et al., 2014) and an update of all reaction coefficients and absorption cross-sections according to Sander et al. (2011). Recently, Sukhodolov et al. (2016) showed that the SOCOL photolysis scheme is in good agreement with high-resolution radiative transfer codes in the stratosphere, in terms of representing both the climate state and the solar cycle. Ball et al. (2016a) demonstrated good agreement between modeled ozone and observations over the 11-year solar cycle. These results provide evidence that the CCM SOCOL is an effective tool to simulate solar effects in the middle atmosphere.

We performed two experiments for years 2003-2005: the reference 30-member ensemble run with daily variable irradiance fluxes (RefSun) and a 30-member ensemble run with irradiance fluxes intra-seasonally smoothed (NoSun). Both SSI forcings are illustrated in Figure 1. Other boundary conditions mostly follow the emission scenario REF-C1 from the IGAC/SPARC Chemistry-Climate Model Initiative (CCMI; Eyring et al., 2013), which includes prescribed sea surface temperatures and sea ice coverages; initial conditions are taken from a CCMI model simulation for 1960-2012. Ensemble members differ by slight changes in CO₂ concentration.

2.4 Observations

Since the ozone response to SSI variability on the solar rotational time scale has been well established (Bossay et al., 2015 and reference therein), we focus here only on the analysis of the observed temperature. We analyze the newly recalibrated and merged observations from the Stratospheric Sounding Unit (SSU) instrument onboard NOAA historical polar orbiting satellites. The data are available for 1979-2006 with five-day temporal resolution, but only for 3 SSU channels corresponding to 15, 5, and 1.5 hPa (~28, 36, 45 km). A detailed description of this dataset is given by Zou et al. (2014). We also exploited the most recent reanalysis datasets ERA-Interim (Dee et al., 2011) and MERRA2 (Molod et al., 2015). Both reanalysis datasets provide daily data with full spatial coverage for a period since 1979. Reanalysis is a unique approach aimed at combining all existing observations. However, in order to fill the gaps in observations, atmospheric models are used, which generally don't apply daily variable solar forcing. Therefore, the effect of short-term solar variability would be damped or absent through smoothing of the pre-prepared input irradiance data. In this context, comparison of direct observations and reanalysis data is interesting.

2.5 Methods of analysis

Various statistical methods may be used to identify the solar rotational signal in the middle atmosphere. Here we demonstrate that, even following a classical procedure (e.g. Hood and Zhou, 1998), it is possible to obtain a robust response. To remove annual, semi-annual and long-term variability, as well as short-term noise we apply 35-day and 7-day boxcar smoothing. We have compared this approach to other methods, such as Butterworth and Gaussian bandpass filters, and found no significant difference. In Section 3.1 we use filtered data to calculate lagged correlations of the simulated temperature response with solar irradiance at 205 nm (cross-correlation). We calculate the statistical significance of the obtained correlation coefficients in a similar way to Hood (2016), i.e. taking into account autocorrelation of 7-day averages. We used the lowest effective sample size found from analysis of different variables (temperature from different sources, ozone, hydroxyl, vertical residual circulation) at different locations and estimated that for the 1, 2, and 3-year periods, used here, correlation coefficients higher than 0.3, 0.2, and 0.17, respectively, are statistically significant at the 95% confidence level. In Section 3.2 we analyze sensitivities of atmospheric parameters to changes in irradiance fluxes derived by a linear regression technique taking into account autocorrelation of residuals. Gruzdev et al. (2009) found similar results comparing ozone and temperature sensitivities estimated using linear regression and sensitivities estimated using cross-spectral analysis. Finally, in Section 3.3 we apply continuous

wavelet transforms based on the Morlet wavelet (Torrence and Compo, 1998) in order to analyze and compare amplitudes and periods of variations of 3-year and 30-year time series of irradiance and stratospheric temperature.

3. Results

3.1 Cross-correlation analysis

Figure 2 represents the lagged correlation between solar irradiance at 205 nm (I205) and tropical hydroxyl, ozone, and temperature calculated using the 1-D RCPM model. Since the 1-D model doesn't include any dynamics except diffusion and convection, the presented results are mostly defined by the chemical and radiative processes. This explains the high correlation coefficients (> 0.9), not only for short-lived hydroxyl, but also for ozone and temperature. Hydroxyl and ozone responses are consistent with, and well explained by, earlier studies (Rozanov et al., 2006; Shapiro et al., 2011). Hydroxyl is a short-lived species produced by water vapor, oxygen and ozone photolysis mostly by the highly variable shortwave (< 320 nm) solar irradiance. Although the hydroxyl response is pronounced throughout the entire middle atmosphere, its negative effect on ozone enhances in the lower mesosphere, where water vapor content is substantial and air density is low. This effect is illustrated by the negative ozone correlation at a zero phase lag around 70 km. The zero-lag hydroxyl response also contributes to the negative lag of the ozone response in the upper stratosphere (Hood 1986). In the stratosphere, ozone is positively correlated with irradiance. Starting from 40 km and moving downwards, phase lags of the ozone and hydroxyl responses gradually increase from slightly negative up to 5 and 18 days, respectively, both due to the increasing air density and due to the decreasing of UV-flux related to the increased absorption of additionally produced ozone in the upper and middle stratosphere. The solar signal in temperature is lagged by several days around the stratopause region. From 45 to 20 km the phase lag of temperature response also gradually increases up to 10 days. Around 70 km, a temperature phase lag increases by several days, likely due to the ozone destruction by additional HO_x .

However, all previous 3-D modeling studies could not identify such a pronounced correlation of the middle atmospheric temperature with irradiance, neither with a realistic solar forcing (Rozanov et al., 2006; Kubin et al., 2011), nor with an artificial sinusoidal signal and, thus, stable in time (Gruzdev et al., 2009). The failure to confirm a solar rotational signal is true for observational studies, which led to a general conclusion that it is basically impossible because of time series contamination due to internal atmospheric variability. Some studies (Hood and Zhou, 1998; Williams et al., 2001; Rozanov et al., 2006) even reported a negative lag in the upper stratosphere temperature response, the region of highest heating, which indicates that the correlation can be both coincidentally decreased and also increased. Gruzdev et al, (2009) also showed higher correlation in a simulation with a constant solar forcing than in one with the forcing changing in time. The detection of an ozone response has similar problems, though to a lesser extent due to a generally shorter relaxation time.

(Figure 2)

Internal short-term variations in the middle atmosphere are driven by changes in atmospheric waves, their non-linear interaction with the zonal mean-flow, and with each other (Scaife and James, 2001); they can also be produced within the stratosphere itself (Scott and Polvani, 2006). On intra-seasonal timescales these processes are not well reproduced in global models compared to observations because of uncertainties in initial and boundary conditions, low resolution and other simplifications. On the one hand, this complicates the direct comparison of observations and models on daily timescales especially in the middle atmosphere. On the other hand, the stochastic nature of short-term dynamical fluctuations in models should diminish their impact on the mean state if a sufficient number of model realizations are considered, even if they are not normally distributed around a mean state. Following this assumption, we performed 30-member ensemble experiments with the 3-D CCM, both with and without rotational variability in the input irradiance fluxes. Figure 3 illustrates the 3-year results for hydroxyl, ozone and temperature obtained with the 3-D CCM SOCOLv3. These results mostly replicate the 1-D results presented in Figure 2 in terms of phase lag behavior. Correlation coefficients are generally lower especially for temperature, likely due to some remaining contamination by random internal variability as well as by dynamical feedbacks following thermal wind balance, which is omitted in the 1-D model. In the mesosphere, 3-D modeling results also don't reveal a local increase of the temperature signal phase lag as in the 1-D model.

(Figure 3)

The period considered (2003-2005) contains short periods of weak rotational cycles between impulsive events of active region emergence that produce larger amplitude 'clean' solar rotational variability of about six-ten-months (Woods et al., 2015). Additionally, there is a decrease in solar activity as the Sun heads towards solar minimum. In order to investigate the stationarity of the results we performed a cross-correlation analysis for one-year-long moving windows with 90-day steps. The window length chosen allows seasonal variance to be averaged out and to track differences imposed by changes in the variability of solar irradiance. The temperature results are presented in Figure 4 and show the expected higher correlations than by considering the whole 3-year period. Phase lag responses remain quite stable over the first four windows; the next three windows reveal degradation of the signal in the lower stratosphere, where the importance of dynamical feedbacks is larger. The last two windows reveal lower correlations and blurred time lag patterns. Thus, it is highly suggestive that the temperature response at the beginning of the analyzed period is solar variability related, since its evolution follows the evolution of the irradiance signal. This is supported by the same temporal analysis of the signal in ozone (Figure S1 in supplementary materials (SM)), which also shows a similar, though less pronounced, degradation of the signal with time. The amplitude of the rotational cycle drops by about a factor of two in three years, but the cycle is still well-pronounced at the end of the period considered. This implies that there may be a certain threshold in signal-to-noise ratio, after which it is difficult to detect a robust signal in temperature even using our method. Based on this, we use only the first two years instead of three for the cross-correlation analysis and calculation of sensitivity coefficients that follows. Analysis of separate ensemble members revealed that in some cases the results are similar to the ensemble mean (Figure S2 in SM) while in other cases the modulation by dynamics can be so high that the response to the solar signal is not pronounced even at the beginning of the considered period

(Figure S3 in SM).

(Figure 4)

The two-year cross correlation functions of the RefSun experiment, NoSun experiment, reanalysis data, and NOAA SSU temperature observations are presented in Figure 5. Compared to Figure 3, the correlation coefficients obtained from RefSun are now more pronounced and maximize around 50 km at a 0.5 level. The ensemble mean temperature response in the absence of solar rotational forcing (NoSun) doesn't exceed a correlation of 0.17, which supports our findings. The temperature variation in ERA-Interim and MERRA2 reanalysis data are similar to each other. However, reanalysis signals are vague and not consistent with our 1-D and 3-D modeling results. The only consistent feature is a positive correlation (up to 0.34) near the stratopause with slightly positive lag. NOAA SSU also reveals quite low correlation and no lag in the stratosphere. Similar results are shown in Figure 6 as a function of latitude for tropical and subtropical regions. Correlation coefficients derived from the ensemble mean of the NoSun run barely exceed 0.2. RefSun results reveal correlations up to 0.62 around the stratopause. Both reanalysis datasets reveal generally lower correlations, although also peaking around the tropical stratopause. Another peak in the correlation is found in the lower stratosphere, although a temporal analysis (as in Figure 4) showed that this peak is rather occasional. NOAA SSU observations cover a region between these two peaks and are generally consistent with reanalysis. We also plot the correlation obtained from one of the ensemble members from the NoSun experiment, in order to demonstrate how variable and coincidental a single model realization can be even without solar rotational forcing. Even though the lag pattern is different to what is shown for the RefSun results and theoretically expected, the correlation can exceed 0.4.

(Figure 5)

(Figure 6)

As was pointed out in earlier studies (e.g. Rozanov et al., 2006; Gruzdev et al., 2009), the atmosphere produces variability at a broad range of frequencies including variations with periods close to 27-days. When analyzing the time series as they are, it is hard to distinguish between different processes, since, unlike for longer timescales (Kuchar et al., 2015), so far there is no defined proxy for short-term middle atmospheric variability except solar irradiance. Detailed analysis of separate ensemble members of the NoSun experiment showed that the amplitude of random oscillations can be several times larger than the amplitude of variations given by the ensemble mean RefSun run. In order to estimate these amplitudes, in the next section we analyze sensitivities (i.e. the response of atmospheric parameters to a certain change in forcing) derived from our experiments and observational data.

3.2 Sensitivity analysis

Figure 7 presents sensitivities of tropical mean hydroxyl, ozone and temperature to a 1% change in irradiance at 205 nm (I205), calculated at the time lag of maximum positive correlation as obtained in

Figures 3 and 5. Note that ozone and hydroxyl responses to irradiance change are actually negative above ~ 69 km and below ~ 38 km, respectively. Thin lines represent results obtained from separate ensemble members, while thick lines represent ensemble means, temperature reanalysis and observations. Dashed lines indicate cases when the related correlation is lower than 0.2 and therefore not statistically significant. Calculated hydroxyl sensitivity maximizes around 70 km at a level of 0.7%. This is consistent with 3-D modeling results of Gruzdev et al. (2009), but several times lower than obtained by Shapiro et al. (2012) from MLS Aura observations. Mesospheric photochemistry of water vapor is very sensitive to solar irradiance changes but remains the most problematic part of radiative transfer in current global models, since very simplified parameterizations are used to calculate the photolysis of H_2O and O_2 by the solar Lyman- α line, and this is also an important issue of the SOCOL photolysis scheme (Sukhodolov et al., 2016). Therefore, more detailed analysis is needed, also because Shapiro et al. (2012) is the only observational study performed so far in this field. Ozone sensitivity is generally consistent with all previous modeling and observational estimates. We obtain a maximum ozone sensitivity of 0.44% at around 40 km. Another large sensitivity maximum is found in the mesosphere, where the ozone concentration is small but more irradiance is available. The modeled stratospheric temperature sensitivity maximizes around ~ 50 km at a level of about 0.1 K per 1 % change of I205. In the mesosphere this value increases to 0.21 K.

(Figure 7)

Results of separate ensemble members with and without solar rotational forcing reveal a high spread of sensitivities, which is the most pronounced for temperature. This highlights the importance of internal variability of the atmosphere in modulating the regression results, because even for quite high signal-to-noise ratio cases (e.g. ozone in the upper stratosphere) ensemble members can differ substantially. When we reduce the noise by calculating the ensemble means, NoSun and RefSun cases become clearly different, which allows us to conclude the solar origin of the derived RefSun sensitivities of hydroxyl, ozone, and temperature. However, this is not the case in the lower stratosphere (below ~ 30 km), where results of RefSun and NoSun ensemble mean cases overlap. Given also the low and thus not statistically significant correlations in this region, we can conclude that the lower stratospheric solar signal in hydroxyl, ozone, and temperature is not distinguishable from internal variability in our experiment even with 30 ensemble members.

Temperature sensitivities calculated from reanalysis data are higher than in our RefSun ensemble mean case in the upper and lower stratosphere. This is more consistent with previous estimates derived using linear regression (Gruzdev et al., 2009 and references therein). However, at most altitudes reanalysis and NOAA SSU sensitivities are not supported by a pronounced correlation. Moreover, observations or reanalysis data can be considered an analogue to a single model realization, which, as we showed above, is subject to a high modulation by dynamics. Therefore, even when the related correlation is statistically significant, the presented reanalysis and NOAA SSU sensitivities are rather doubtful. We don't plot confidence intervals for the sake of better visibility, because due to low effective sample size (96) and autocorrelation of residuals the fitting 2-sigma uncertainty in many cases exceeds the plotting domain.

3.3 Wavelet analysis

Figure 8 shows wavelet analysis of a 3-year period of I205 and zonal mean tropical mean (21S-21N) temperature anomalies at ~45 km derived from two modeling experiments, two reanalysis datasets, and the observational dataset. This analysis further strengthens the main conclusion that the solar rotational signal is strongly modulated by internal atmospheric variations by showing that the ensemble mean RefSun temperature shows the closest agreement to the I205 global power spectrum and its temporal distribution. Era-Interim and MERRA2 datasets show a peak periodicity close to 27 days, although both datasets reveal a strong importance of longer periodicities. Longer period signals also dominate in NOAA SSU data. Results of the experiment with the constant solar forcing (NoSun) reveal a quite stable signal of a broad 25-40 days periodicity range, which should be fully attributed to the internal variability of the atmospheric system. The same is true for other vertical levels of the stratosphere above ~30 km: the rotational signal is the clearest in the RefSun ensemble mean data, while all other time series are more contaminated by lower frequency variability. A pronounced 27-day signal in reanalysis temperature is only visible around 50 km, which appears to be the only region revealing a high enough signal-to-noise ratio for the considered period of time, as it is also the case in cross-correlation results (Fig. 5).

(Figure 8)

By means of continuous wavelet transform, we then analyzed the randomness in time of the intra-seasonal temperature variability in reanalysis data. Figure 9 shows time-resolved power spectra densities for ERA-Interim reanalysis zonal mean tropical mean temperature anomalies at 48 km and solar irradiance at 205 nm for 1980-2007 years period. This period covers 3 maxima of the 11-year cycle, which is clearly pronounced in irradiance spectral densities by increases of the signal always peaking at the 27-day period. Temperature analysis reveals pronounced episodic variations with higher periods appearing roughly twice per year. Increases of spectra density around 27-day periods don't systematically follow similar changes in the irradiance signal and in most cases they have low spectral density. Thus, during solar maxima there can be no pronounced increase of rotational variability in temperature and vice-versa for solar minima. Long-term wavelet analysis of MERRA2 and NOAA SSU data revealed similar results (not shown). We discuss the origin of this inconsistency in the following section.

(Figure 9)

4. Conclusions and discussion

We performed an ensemble modeling experiment with the CCM SOCOL in order to study the middle atmosphere response to solar rotational variability. To derive a rotational signal from time-series of atmospheric variables we used standard statistical methods such as bandpass filtering, cross-

correlation, linear regression and wavelet transform. For the first time, we show a well pronounced temperature response in the tropical middle atmosphere to solar irradiance variability on a solar rotation time scale. Phase lag patterns of temperature and ozone responses are found to be in agreement with our results from 1-D modeling (also performed and analyzed here); the 1-D modeling was used to evaluate the atmospheric response in the absence of dynamics and internal variability. A robust correlation of irradiance with temperature and ozone suggests that our sensitivity profiles obtained through linear regression are also robust. In the stratosphere, we report a maximum temperature sensitivity around the stratopause of about 0.1 K per 1% change of irradiance at 205 nm and ozone sensitivity peaking around 40 km at a level of ~ 0.44 % per 1% change of irradiance at 205 nm. These ozone sensitivities appear to be in agreement with the latest estimates of the 11-year solar cycle (Ball et al., 2016a), while temperature sensitivities appear to be at the lowest boundary of the previously reported values (Mitchell et al., 2014; 2015). Therefore our current results suggest a lower warming of the tropical stratosphere than previously estimated, and thus a lower forcing for the initiation of the “top-down” mechanism (Haigh, 1996; Kodera and Kuroda, 2002) or its higher sensitivity to the temperature changes in the upper stratosphere. With respect to the 11-year cycle and “top-down” effect, our results also suggest that the lower stratospheric signal is highly uncertain, but likely there is no lower stratospheric peak of response, both in ozone and temperature, because we don’t obtain an increase of sensitivity at these altitudes even in separate ensemble members, temperature reanalysis data and observations. Hood (2003) previously reported a clear peak temperature response at ~ 100 hPa in NCEP-NCAR reanalysis daily data. However, in agreement with our conclusion, Hood (2016) recently performed a similar analysis to Hood (2003) with updated NCEP-NCAR and Era-Interim data and found that there was a data processing problem in the old NCEP-NCAR reanalysis, which was responsible for the enhancement of the response in the lower stratosphere.

An analysis of separate ensemble members, a run without rotational variability, and temperature reanalysis and observations all showed that the internal variability of the atmosphere dominates the variability of the stratospheric temperature on solar rotation time scales. This makes an extraction of the 27-day signal from observations and single model realizations using standard techniques difficult. Ozone and hydroxyl time-series undergo the same problems, though to a lesser extent due to a lower relaxation time defined by the photochemical lifetime changing with height. One way to beat the internal variability is to choose years under solar maximum conditions, thus, increasing the signal-to-noise ratio. Recently, Hood (2016) used similar statistical methods to analyze Era-Interim temperature data and reported traces of the 27-day variability in the lower stratospheric temperature. Although the reported correlation coefficients didn’t exceed 0.21, they were statistically significant due to a long time period analyzed (9 years of solar maximum conditions). Ensemble modeling allows for further increasing of the signal-to-noise ratio. To illustrate this effect we performed a bootstrapping procedure by calculating cross-correlation functions of ensemble mean tropical ozone and temperature with irradiance (as in Figure 5) for the ensemble sizes changing from 1 to 30 and randomly sampled 100 times from the calculated 30 members. The results are shown in Figure 10 for 3 atmospheric levels representing the higher, the middle, and the lower stratosphere. As one can see, the increase of the ensemble size allows for a significant increase of correlations and decrease of uncertainty in the upper stratosphere for ozone and temperature. In the middle stratosphere the effect is still well pronounced

but not so dramatic, while in the lower stratosphere the ensemble size increase allows only for a reduction in the correlation uncertainty without increasing the correlation coefficient itself. Therefore for the lower stratospheric temperature we are in agreement with Hood's value of ~ 0.2 . A low but statistically significant correlation implies that at daily timescales we understand only a small part of variance. At the upper atmospheric levels of the ensemble mean case we can obtain a maximum correlation of ~ 0.6 for temperature around the stratopause (Figures 4, 6), which is the highest value reported so far. However, this means that still at most we are able to explain only $\sim 36\%$ of the tropical temperature daily variance, while the rest can be attributed to noise, internal variability, or uncertainty. There were other attempts to perform quasi-ensemble simulations (Rozanov et al., 2006; Gruzdev et al. 2009; Grenfell et al., 2010; Kubin et al., 2011;), however they all were of about 1-10 realizations of 1 year, while in our case for the final estimates we used 30 realizations of 2 years of data. From Figure 10 it becomes clear that 10 years (i.e. 5 ensemble members) helps to reduce only about a half of the uncertainty compared to 60 years (30 members) and there is still some potential for further improvement.

Our wavelet transform results also revealed a high importance of the internal variability. Significant increase of spectral density during southern and northern wintertimes even in the absence of the solar rotational forcing suggests that such occurrence can be likely related to the evolution of the polar stratospheric vortices as parts of the Brewer-Dobson Circulation (BDC) system driven by atmospheric waves. Such variability has already been discussed before (Holton and Mass, 1976; Chandra et al., 1986, Kuroda and Kodera, 2001) but has never been used as a regressor to account for the internal variability in the stratosphere on daily timescales. At longer timescales, besides the six or seven regressors usually used to describe the variability of ozone and temperature in the stratosphere (e.g. Mitchell et al., 2014; Kuchar et al., 2015), there were several attempts to decrease the remaining uncertainty by producing an additional proxy, which would describe the non-tropical wave-forcing. The latest study by Ball et al. (2016b) confirmed a clear connection between wintertime high-latitude and tropical regions of the stratosphere and mesosphere on a monthly timescale and formulated a proxy based on mid-latitude stratospheric temperatures, which enabled the uncertainty in the tropical upper stratosphere to be reduced by up to 30% and 25% in temperature and ozone, respectively. We applied the approach of Ball et al. (2016b) to calculate a daily proxy (2.5–6.3 hPa 43–49S temperature for May–October and 4–10 hPa 52–57N temperature for November–April). This proxy was found to correlate with tropical temperature by up to 0.8-0.85 in the upper stratosphere for the long 30 year period of ERA-Interim data (Figure S4 in SM) and for the 3-year period used here for modeling. As in Chandra et al. (1986), this proxy reveals a wintertime periodicity of 3-7 weeks more pronounced in the northern hemisphere, a hemisphere where wave-activity originating in the troposphere is more prevalent.

Initial tests employing this index showed that at some pressure levels of zonal mean tropical mean ensemble mean data, while separately the index and irradiance correlations with temperature do not exceed ~ 0.45 , the inclusion of both these variables as independent regressors can reveal the correlation of up to 0.7. However, although the proxy seems not to be correlated with irradiance, its independence as a regressor is questionable; following the solar “top-down” hypothesis the polar regions should also

be affected by the solar forcing that starts with a warming of the higher stratosphere. In support of this, our analysis of the cross-correlation functions of irradiance and vertical residual circulation in the tropical stratosphere (Figure S5 in SM) showed that, even though the correlation coefficients are not high, in the middle stratosphere they generally replicate the cross-correlation pattern of temperature (Figure 4) but with an opposite sign, suggesting that the BDC responds to temperature changes in the upper stratosphere. A similar connection was shown by Grenfell et al. (2010). This is in agreement with the idea that the solar-induced warming of the upper stratosphere should cause a slowing of the BDC (Haigh, 1996; Kodera and Kuroda, 2002), but note, again, that in our case we do not report any peak of warming in the lower stratosphere due to less adiabatic cooling. Nevertheless, a careful development and inclusion of a similar proxy for intraseasonal dynamical variability should be a natural next step in deriving the true solar rotational signal and improving our understanding of the middle atmospheric variability. The possible problems related to aliasing should be then also studied.

Recently, superposed epoch analysis was used to derive the rotational signal in temperature observations in the mesosphere (Thomas et al., 2015) and stratosphere (Garfinkel et al., 2015). This approach seems to be closest to ensemble modeling in the context of a strong irradiance signal influence from internal variations, because it performs an analysis similar to ensemble modeling picking particular cycles or members fulfilling certain criteria related to forcing. A study incorporating both this statistical approach and the ensemble modeling approach we employ here would be useful to confirm if the solar rotational signal in the atmosphere is at all detectable from single-member model simulations.

Friedrich et al. (2014) and Fytterer et al. (2015) highlighted a potential effect from solar modulated geomagnetic activity by showing pronounced 27-day variability in nighttime O_3 , OH , and NO_2 in the mesosphere and upper stratosphere. At 11-year solar cycle time scales, Arsenovic et al. (2016) also showed that the polar vortex could be modulated by middle range energy electron precipitation. Even though there is no direct effect at tropical latitudes, as there are almost no energetic particles entering the atmosphere there, ozone variations in the polar regions can potentially affect the evolution of the polar vortices and, thus, the BDC. The most distinct surface response to the 11-year solar cycle appears some years after the 11-year maximum (e.g. Gray et al., 2013), which can also hint to the important role of magnetospheric energetic particle effects (Maliniemi et al., 2014), since this forcing is also shifted in phase by a couple of years compared to the solar UV forcing. On daily timescales, particles may then also modulate the temperature proxy we discussed above, though this is speculative. Therefore a modeling study considering both solar UV and energetic particle effects on daily timescales would be of interest in this context.

Acknowledgments

This research was funded in part by the Swiss National Science Foundation under grant agreements 200020_140573, 200020_153302, 200020_149182 (SILA), 200020_163206 (SIMA), CRSI122-130642 (FUPSOL) and CRSI12-14765920 (FUPSOL II). Fruitful discussions with the COST Action ES1005 TOSCA (<http://www.tosca-cost.eu>) community are much appreciated. Wavelet software was provided by C. Torrence and G. Compo and is available at URL:

<http://paos.colorado.edu/research/wavelets/>.

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Fig. 1. Variability of the applied solar irradiance at 205 nm (%). Black – original daily NRL SSI data (RefSun); blue – NRL SSI data with the short-term variability removed (NoSun).

Fig. 2. Cross-correlation function of the tropical hydroxyl (left panel), ozone (central panel), and temperature (right panel) versus the solar irradiance at 205 nm, obtained using the 1-D RCPM model for 2003-2005 period. Contour lines are plotted at 0.1 intervals; red represents positive correlations, blue negative.

Fig. 3. Cross-correlation function of ensemble mean tropical mean (21S-21N) zonal mean hydroxyl (left panel), ozone (central panel), temperature (right panel) versus the solar irradiance at 205 nm, obtained using the 3-D SOCOLv3 model for the 2003-2005 years period. Contour lines are plotted at 0.1 intervals; red represents positive correlations, blue negative.

Fig. 4. Cross-correlation functions of ensemble mean tropical mean (21S-21N) zonal mean temperature versus the solar irradiance at 205 nm, obtained using the 3-D SOCOLv3 model for 1-year period windows moving with 90 day steps. Contour lines are plotted at 0.1 intervals; red represents positive correlations, blue negative.

Fig. 5. Cross-correlation functions of tropical mean (21S-21N) zonal mean temperature versus the solar irradiance at 205 nm, obtained from the ensemble mean NoSun run), ensemble mean RefSun run, ERA-Interim and MERRA2 reanalysis, and NOAA SSU observations for the 2003-2004 years period. Contour lines are plotted at 0.1 intervals; red represents positive correlations, blue negative.

Fig. 6. Correlation fields of zonal mean temperature versus the solar irradiance at 205 nm as a function of latitude obtained from NoSun ensemble mean, NoSun 21st ensemble member, RefSun ensemble mean, MERRA2 and ERA-Interim reanalysis, and NOAA SSU observations for the 2003-2004 years period. Correlation coefficients are calculated as maximum absolute correlation at all lags. Contour lines are plotted with 0.1 steps. Areas with a correlation coefficient higher than 0.2 are colored.

Fig. 7. Sensitivities of tropical mean hydroxyl, ozone and temperature to 1% change in I205. Sensitivities are derived using a linear regression method at a lag of maximum positive correlation. Thin lines represent results obtained from separate ensemble members; thick lines represent results obtained from ensemble mean cases, reanalysis and observations. Dashed lines represent cases when the related correlation coefficient is lower than 0.2.

Fig. 8. Time-resolved normalized power spectra densities (left panels) estimated from continuous wavelet transform (CWT) of irradiance at 205 nm (I205), and temperature at ~45 km derived from the NoSun and RefSun experiments (ensemble mean), from Era-Interim and MERRA2 reanalysis data, and NOAA SSU observations. The dotted contour lines represent the 90% confidence level. Green hatched areas in the upper corners of each panel represent the “cone-of-influence”, anything below these areas is doubtful. Black thin lines represent period of 27 days. Right panels: time average of the left panels.

Fig. 9. Time-resolved power spectra densities (scalogram) estimated from continuous wavelet transform (CWT) of irradiance at 205 nm (upper subpanels) and ERA-Interim temperature at 48 km (lower subpanels) for the 1980–2007 years period. Each of the 5 presented panels corresponds to 2000-day periods, to which CWT was applied separately. The dotted contour lines represent the 60% confidence level. Green hatched areas in the upper corners of each panel represent the “cone-of-influence”, anything within these areas is doubtful. Black horizontal lines represent period of 27 days.

Fig. 10. Correlation coefficients of the tropical mean zonal mean ozone and temperature versus I205 calculated as a function of ensemble size. Each ensemble size was randomly sampled from 30 model realizations. Different colors designate different vertical levels. Dashed lines indicate the 2- σ uncertainty revealed by random sampling.