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The Response of the Middle Atmosphere to Long-Term and Short-Term Solar Variability: A Two-Dimensional Model

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A two-dimensional chemical-dynamical-radiative model of the middle atmosphere is used to investigate the potential changes of temperature, ozone, and other chemical constituents in response to variations in the solar ultraviolet flux, associated with the solar rotation (27 days) and the solar cycle (11 years). The model reproduces satisfactorily the response (amplitude and phase) to the 27-day forcing of ozone and temperature in the stratosphere but does not properly explain the ozone and temperature responses of opposite sign observed near 70 km altitude. The change in the ozone column abundance associated with the 27-day solar forcing is estimated to be less than 0.5%. Variations in middle and upper atmospheric ozone concentrations and temperatures induced by solar variations on the 11-year time scale are not negligible compared to changes produced as a result of human activities over the same period of time. The calculated change in the ozone column abundance from solar minimum to solar maximum conditions is of the order of 1.1–1.3% in the tropics and increases with latitude, especially in winter, to reach up to 1.5–1.7% in the polar regions. There are large unexplained differences between the calculated and observed stratospheric responses. For example, in the photochemically controlled region of the upper stratosphere, the model seems to underestimate the ozone response by more than a factor 2. In addition, the negative temperature and ozone responses observed in the lower stratosphere cannot be reproduced by the model. Potential dynamical feedbacks exist, but reliable data sets covering a period longer than one solar cycle have to become available before this problem can adequately be addressed.

1. INTRODUCTION

It has been recognized for a long time that the electromagnetic flux emitted by the Sun is not perfectly constant in time, but that it slightly varies, especially at shortest wavelengths, on different time scales (see, e.g., *Lean* [1987], *Donnelly* [1988, 1991], *Simon* [1989], *Simon and Tobiska* [1991] for reviews of the subject). Generally, the most pronounced variability is associated with the 11-year solar cycle. In addition, an observed short-term variation in the solar flux results from the uneven distribution of active regions on the Sun which rotates with an apparent period of 27 days. Figure 1 shows the temporal evolution of the solar flux at Lyman-α (121.6 nm), in the region of the O_2 Schumann-Runge bands (180–200 nm) and in the region of the O_2 Herzberg continuum (205 nm), deduced from the observations by the Solar Mesosphere Explorer (SME) between 1982 and 1988 (P. C. Simon and G. Rottman, personal communication, 1989). The solar flux at Lyman-α affects mainly the atmosphere above 75 km altitude, while the flux at 180–200 nm is mainly absorbed by molecular oxygen in the mesosphere. The solar radiation between 200 and 240 nm is primarily responsible for the formation of ozone in the stratosphere and hence the flux at the particular wavelength of 205 nm is often used as an index to represent the intensity of the solar forcing on the stratosphere. As suggested by Figure 1, the variation between the early 1980s (solar maximum condition) and the mid 1980s (solar minimum condition) is of the order of 50–60% at Lyman-α, 5% at 180–200 nm and 6–7% at 205 nm. The variability over the full wavelength interval ranging between 120 and 300 nm, as estimated by *Rottman* [1988], is shown in Figure 2*a*. These values remain somewhat uncertain because of potential drift in the solar detectors over the 7-year period. Note also that the SME values cover only a fraction of an 11-year solar cycle. Shown in Figure 2*b* is the 27-day solar variability derived for two distinct periods of the solar cycle (active Sun in 1982, and quiet Sun in 1985, respectively) and covering the 120-300 nm spectral region. As suggested by Figures 2*a* and 2*b*, the variability decreases substantially with increasing wavelength. In addition, a sharp drop in the solar variation is visible at 208 nm (aluminum edge in the solar spectrum) and a relatively large amplitude in the solar signal is noticeable at 280 nm (magnesium II).

Earlier theoretical studies [*Frederick*, 1977; *Callis and Nealy*, 1978; *Penner and Chang*, 1978; *Brasseur and Simon*, 1981; *Garcia et al.*, 1984; *Callis et al.*, 1985; *Eckman*, 1986; *Brasseur et al.*, 1987; *Wuebbles et al.*, 1991] have suggested that periodic changes in the solar flux associated with the solar cycle and the rotation of the Sun should affect the chemical composition and thermal structure of the stratosphere and the mesosphere. Because reliable measurement records for the middle and upper stratosphere span only slightly more than one solar cycle, statistical evidence for the response of the middle atmosphere to the 11-year solar variation cannot yet be established with full confidence. Nevertheless, analyses of total ozone column abundances measured by the Nimbus 7 total ozone mapping spectrometer (TOMS) [*Chandra*, 1991; *Stolarski et al.*, 1991; *Hood and McCormack*, 1992], of upper stratospheric ozone densities observed by the Nimbus 7 solar backscattered ultraviolet (SBUV) instrument (L. L. Hood et al., Quasi-decadal variability of the stratosphere: Influence of long-term solar ultraviolet variations, submitted to *Journal of*

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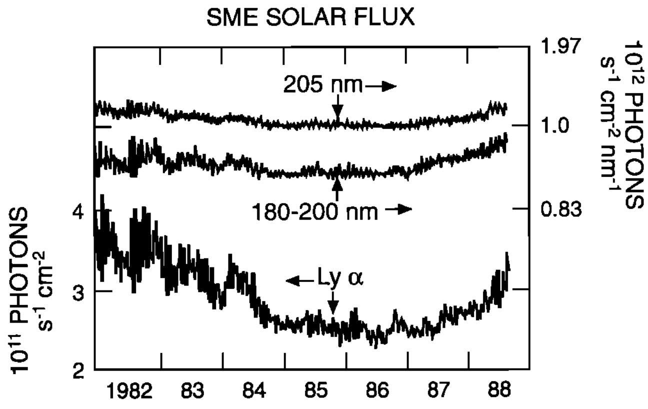


Fig. 1. Solar actinic flux measured by the Solar Mesosphere Explorer at different wavelengths (Lyman-α, 180–200 nm, 205 nm) between 1982 and 1988. (From G. Rottman and P. C. Simon, personal communication, 1990.)

Atmospheric Sciences, 1993) (hereinafter referred to as (Hood et al., submitted manuscript)) and of temperatures archived by the NOAA National Meteorological Center (NMC) (Hood et al., submitted manuscript) have revealed a likely correlation with solar cycle. Data from the Stratospheric Aerosol and Gas Experiment II (SAGE II), obtained over a period of increasing solar activity (after 1985), also suggest a positive relation between ozone in the upper atmosphere and solar ultraviolet flux [*Keating et al.*, 1993*a*]. Observations of stratospheric temperatures measured by radiosondes [*Labitzke et al.*, 1986; *Angell*, 1988, 1991], rocketsondes [*Angell*, 1991] and lidar [*Chanin et al.*, 1987; *Hauchecorne et al.*, 1991], as well as measurements of total ozone by ground-based Dobson instruments [*Angell*, 1989; *World Meteorological Organization*, 1988] also suggest a possible response to the radiative forcing associated with the solar cycle. The effects on the dynamics of the middle atmosphere remain a matter of discussion. The observational evidence of an association between the 11-year solar cycle, the quasi-biennial oscillation (QBO) and the interannual atmospheric variability reported by *Labitzke and van Loon* [1988] was questioned by *Baldwin and Dunkerton* [1989] and by *Salby and Shea* [1991] on statistical grounds and has not yet been properly explained. *Kodera et al.* [1991] showed that the observed temperature anomalies reported in winter by *Labitzke and van Loon* [1988], including their dependence on the phase of the QBO, could be qualitatively reproduced in a general circulation model only if an unrealistic change in the stratospheric heating rate of at least 30% was used to simulate the solar forcing.

More reliable statistical evidence has been obtained concerning the potential response of the middle atmosphere to short-term solar variations. The analyses by *Gille et al.* [1984], *Hood* [1984, 1986], *Keating et al.* [1985], *Chandra* [1986] and *Hood et al.* [1991], based on satellite observations in the 1980s, have documented ozone and temperature perturbations at the 27-day period.

The purpose of this paper is to provide a new theoretical estimate of the potential changes in the chemical composition of the middle atmosphere in response to natural solar variability. The focus will be on ozone in the stratosphere and mesosphere, but changes in temperature which affect the chemical reaction rates will also be considered. Two cases will be examined. First, the response (amplitude and phase) of the middle atmosphere to the 27-day solar forcing will be considered. This part of the work is an extension of the study performed earlier by *Brasseur et al.* [1987] using a simple one-dimensional model. In the present paper a two-dimensional (2-D) model with a formulation of coupled chemical, radiative and dynamical processes accounts for potential feedbacks which could not be simulated by earlier one-dimensional (1-D) models. In a second part of the present study, the response of the atmosphere to long-term changes in the solar forcing will be discussed only briefly, since a recent paper by *Huang and Brasseur* [1993] addresses this question.

The paper is organized in the following way: In section 2 the model used in this study will be described and the adopted variation in the solar actinic flux will be presented. In section 3 the response of the middle atmosphere to the 27-day solar variability will be discussed. Section 4 will present a similar analysis, but for the 11-year forcing. Differences between theoretical predictions and observational estimates will be stressed. Finally, the key conclusions will be summarized in section 5.

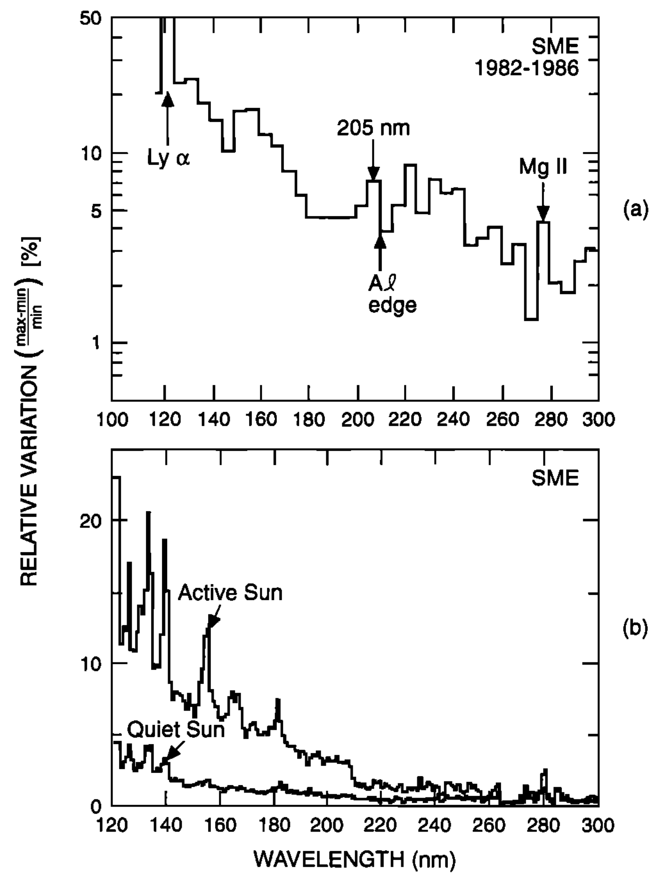


Fig. 2. Variation (percent) in solar ultraviolet flux as a function of wavelength derived from the measurements by the Solar Mesosphere Explorer. (*a*) Variation between solar maximum and solar minimum conditions (11-year cycle); (*b*) variation over a solar rotation period (27 days) for active and quiet Sun, respectively. (From *Rottman* [1988] and G. Rottman and P. C. Simon (personal communication), 1990).

2. Model Description

The model which is used in the present study [see *Brasseur et al.*, 1990] is two-dimensional (latitude, altitude) and extends from pole to pole and from the surface to the mesopause. Its spatial resolution is 5 degrees in latitude and 1 km in altitude. Chemical, dynamical, and radiative processes are treated interactively above 15 km altitude. Below this level, the distributions of the chemical species are calculated with prescribed temperature and winds.

Dynamical equations are expressed using the transformed Eulerian mean formulation (TEM, also called the residual circulation) proposed by *Andrews and McIntyre* [1976]. Momentum deposition and eddy diffusivities are parameterized as a function of the zonal mean wind and temperature fields, according to *Lindzen* [1981] and *Holton* [1982] in the case of gravity waves, and according to *Hitchman and Brasseur* [1988] in the case of the Rossby waves. The solar forcing is calculated through an explicit spectral integration of the energy deposited, while, below 65 km altitude, the infrared forcing is obtained from the radiative transfer code used in the NCAR community climate model (CCM1) [*Kiehl et al.*, 1987]. Above this level, where local thermodynamical equilibrium (LTE) conditions do not apply, rather than using a complex and computationally expensive non-LTE code, a simple Newtonian cooling approach is adopted instead.

The chemical scheme includes approximately 60 species belonging to the oxygen, hydrogen, nitrogen, chlorine and bromine families, and 120 chemical and photochemical reactions. The solar actinic flux at the top of the atmosphere used for the calculation of the photolysis and shortwave heating rates is taken from *Brasseur and Simon* [1981] and is assumed to be representative of solar minimum conditions. For solar maximum conditions, the flux is increased between 120 and 300 nm by the relative values shown as a function of wavelength in Table 1. Because the data deduced from satellite observations remain approximate over the entire spectral domain, and particularly between 160 to 200 nm, where the sensitivity of the two spectrometers on board SME was low (G. Rottman, personal communication, 1993; see also Figure 2*a*), we have adopted a somewhat idealized solar variability and have, for example, ignored rapid fluctuations of this variability as a function of wavelength. In our study the peak-to-peak variability associated with the 11-year solar cycle is 68% at Lyman-α (P. C. Simon, private communication, 1990) and 6.6% at 205 nm. Between 208 nm (Al edge) and 265 nm, the variation adopted in standard calculations is 3%, although the possibility for a variation as high as 5% is considered in one sensitivity test. For calculations dealing with the 27-day variability, the relative variation at all wavelengths is assumed to be half of that shown in Table 1 (standard case). For the sake of simplicity, we have adopted a wavelength dependence of the variability similar to that of the 11-year cycle. The specific case of Lyman-α will be discussed later in the paper. Because the amplitude of the 27-day solar signal evolves as a function of time, the atmospheric response (e.g., ozone and temperature changes) will be expressed relative to a 1% change in peak-to-trough solar variation at 205 nm. The corresponding amplitudes will, in this case, be referred to as "sensitivities."

3. Ozone and Temperature Responses to the 27-Day Solar Variability

The response of ozone and temperature to short-term solar variability is calculated by applying to the solar flux a sinusoidal perturbation with a peak-to-trough amplitude as

TABLE 1. Adopted Variability in the Solar Flux (11-Year Cycle) as a Function of Wavelength (120–300 nm)

Wavelengths, nm	Peak-to-Peak Variation, %	Variation Relative to 205 nm
Lyman-α (121.6)	68.0	10.3
	Schumann-Runge Continuum	
116–141	24.0	3.6
141–149	14.0	2.1
149–159	20.0	3.0
159–170	14.0	2.1
170–175	10.0	1.5
	Schumann-Runge Bands	
175–180	10.0	1.5
180–182	14.0	2.1
182–185	10.0	1.5
185–191	9.0	1.4
191–200	7.6	1.2
200–202	6.6	1.0
	Herzberg Continuum	
202–208	6.6	1.0
208–243	3.0 (5.0)*	0.45 (0.76)*
	Hartley Band	
243–267	3.0 (5.0)*	0.45 (0.76)*
267–270	0.6	0.09
270–278	2.0	0.30
278–282	6.0	0.91
282–303	0.8	0.12

*High value adopted when specified in the text.

specified in section 2. The solution of a linearized equation for an ozone perturbation suggests that if the forced perturbation acts only on the O_2 photolysis rate (ozone production) and if the temperature feedback is ignored, the amplitude of the ozone signal is proportional to

$$\frac{\Delta O_3}{O_3} \propto \frac{\Delta J_{O_2} \tau}{\left(\omega^2 \tau^2 + 1\right)}$$

and the phase (time lag of the response relative to the solar signal) is given by

$$\Phi = \tan^{-1}(\omega\tau)$$

where $\omega = 2\,\pi/T$ (T being the period of the solar forcing) and τ is the chemical lifetime of ozone [e.g., *Hood*, 1986; *Brasseur et al.*, 1987]. Thus under these assumptions the amplitude of the ozone response increases with the period of the solar signal and an extrapolation of a 27-day response to an 11-year response is not straightforward (see section 5). Furthermore, the time lag of the ozone signal relative to the solar signal is necessarily short in the upper stratosphere, where the photochemical lifetime of ozone is small compared to the period of the solar forcing but is expected to increase towards lower altitudes. These conclusions need to be modified somewhat when other important processes are included in the analysis. For example, in the mesosphere above 65 km altitude, the most important forcing is provided by changes in the photolysis of water vapor by Lyman-α, which produces hydroxyl radicals and effectively reduces the concentration of ozone at these levels. In addition, when the effect of temperature feedback is taken into account, the amplitude and the phase of the ozone response depend on the amplitude and phase of the temperature response and vice-versa [*Hood*, 1986]. In the upper stratosphere, where an increase in temperature tends to enhance the ozone destruction rate, the ozone amplitude calculated with temperature feedback included is smaller than if this feedback is ignored. The phase lag is also modified and can even become negative near the stratopause (ozone response leads the solar signal). This behavior of the ozone/temperature system has been discussed by *Hood* [1986], *Brasseur et al.* [1987], and *Keating et al.* [1987]. Because the phase lag of the temperature calculated by the one-dimensional model of *Brasseur et al.* [1987] was approximately a factor 2 smaller than the observed value, the calculated ozone response could not be accurately calculated. The two-dimensional model, which is used in the present study, includes a more elaborate radiative scheme and allows for a number of potential dynamical feedbacks. It is thus useful to revisit this question.

Figure 3 shows as a function of altitude the calculated response of the equatorial temperature to a solar forcing corresponding to a peak-to-trough amplitude of 3.3% at 205 nm. The largest response (approximately 0.37 K) is found at the stratopause, where the phase lag is equal to 4 days. The amplitude of the temperature response decreases towards lower altitudes and becomes insignificant below 30 km altitude. At the same time, the phase lag increases and reaches 6 days at 40 km and 14 days at 30 km.

The response of equatorial ozone, shown in Figure 4, is relatively complex. Below 70 km, a maximum in the ozone response is predicted at approximately 40 km altitude with a value of 1.2% for a solar variation of 3.3% at 205 nm. At this altitude, the ozone response is exactly in phase with the solar forcing. Lower down, the ozone amplitude decreases and at 30 km, it is only 0.4% with a phase lag of 5.5 days. In the upper stratosphere and lower mesosphere, the amplitude of the ozone response decreases and the ozone response leads the solar forcing by 3 days at 50 km and 4.5 days at 60 km. Above 70 km, the amplitude of the ozone variation increases suddenly and reaches 4% at 75 km. It also becomes exactly out of phase with the solar signal. This is the region of the atmosphere

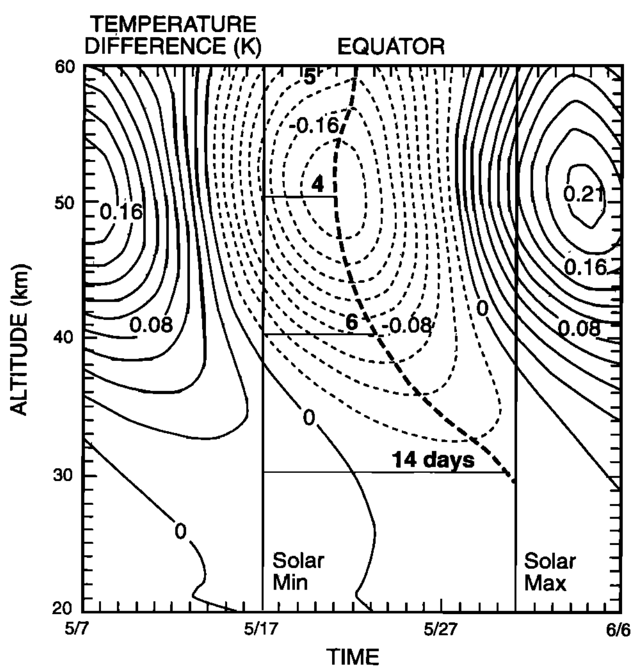


Fig. 3. Calculated response in the atmospheric temperature (Kelvin) to the 27-day variation in solar ultraviolet flux. Values are shown at the equator between 20 and 60 km altitude and cover the period May 7 to June 6. The time lag (in days) of the temperature response to the solar signal is indicated.

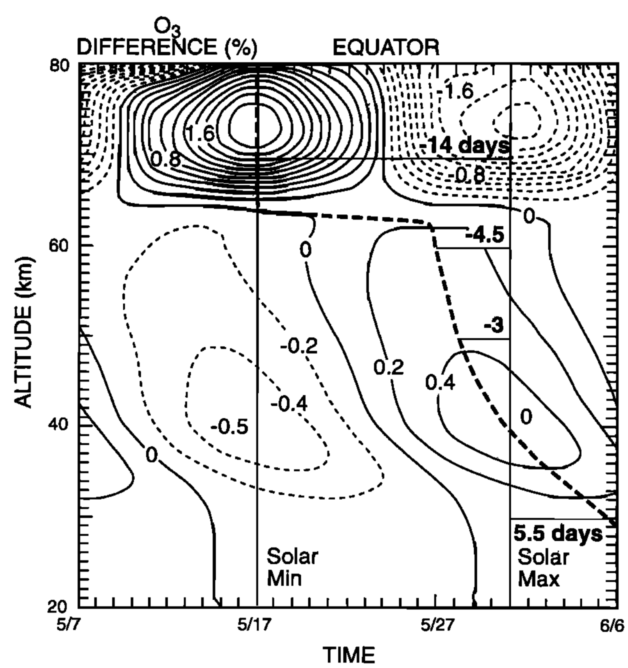
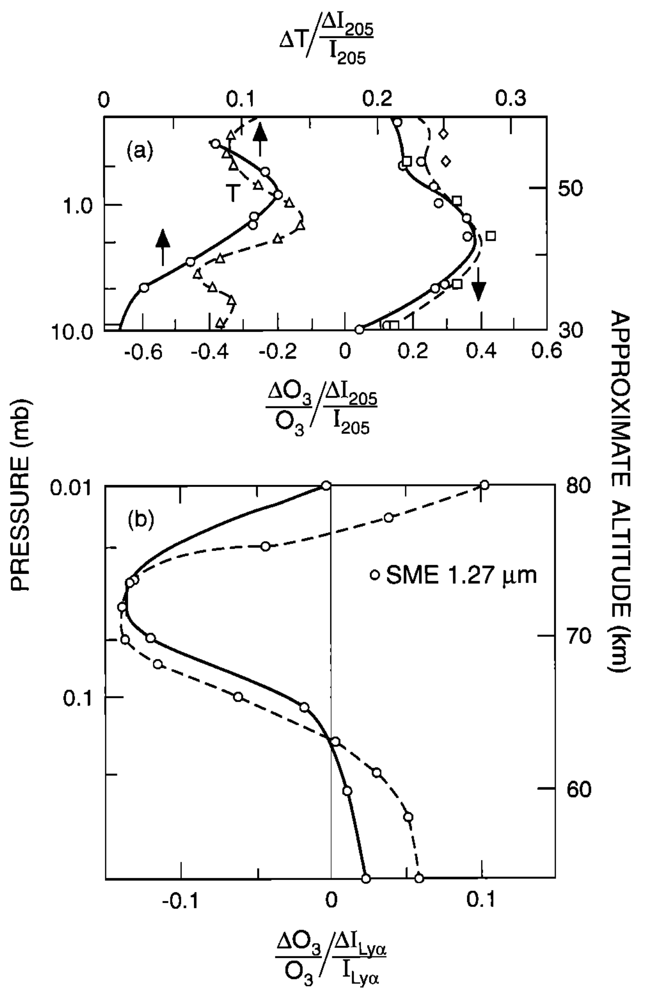


Fig. 4. Same as Figure 3, but for the ozone density (expressed in percent).

where the photochemical effect of the solar flux at Lyman-α becomes important.

In Figures 5*a*-5*d*. the amplitudes (sensitivities) and phases of the ozone and temperature responses calculated at the equator are compared with an analysis of the data averaged over ± 20° latitude, provided by the Nimbus 7 limb infrared monitor of the stratosphere (LIMS), SBUV and stratospheric and mesospheric sounder (SAMS) instruments as well as by the Solar Mesosphere Explorer [*Keating et al.*, 1987; *Hood*, 1986]. The calculated ozone sensitivity is of the order of 0.06 (percent O_3 by percent solar flux at 205 nm) at 30 km, 0.25 at 35 km, 0.38 at 40 km and 45 km, 0.25 at 50 km, 0.20 at 55 km, and 0.15 at 60 km, while the calculated temperature sensitivity is 0.01 K/% at 30 km, 0.06 K/% at 40 km, and 0.12 K/% at 60 km. The overall agreement with the data is good, with a few differences such as in the calculated temperature sensitivity below 40 km (Figure 5*a*). In Figure 5*b*, the ozone sensitivity above 60 km is expressed relative to the solar variability at Lyman-α (as opposed to 205 nm), since the role played by the solar flux in the O_2 Herzberg continuum diminishes above the stratopause, while the influence of Lyman-α becomes dominant. In this case, the ozone sensitivity (relative to Lyman-α) is 0.01 (percent per percent) at 60 km, -0.14 at 70 km and approximately zero at 80 km. Again, the agreement between model and observation is satisfactory. If, however, the ozone response above 65 km is expressed relative to the 205-nm solar radiation, the sensitivity is too large by approximately a factor of 1.5, suggesting that the ratio between the relative solar variation at Lyman-α and at 205 nm is approximately 5–6 rather than 10, as adopted in the present calculation (see Table 1). A ratio of 5 is consistent with the preliminary analysis of solar measurements made by the SOLSTICE instrument on board the Upper Atmosphere Research Satellite (G. J. Rottman, personal communication, 1993).

The phase lags for the temperature (Figure 5*c*) (14 days at 30 km, 7 days at 40 km, 4 days at 50 km, 5 days at 60 km) and ozone density (Figure 5*d*) (4.5 days at 30 km, 0 days at 40 km, -3 days at 50 km) calculated by the 2-D model are in much better agreement with observations than those derived earlier by 1-D models [*Eckman*, 1986; *Brasseur et al.*, 1987]. The difference



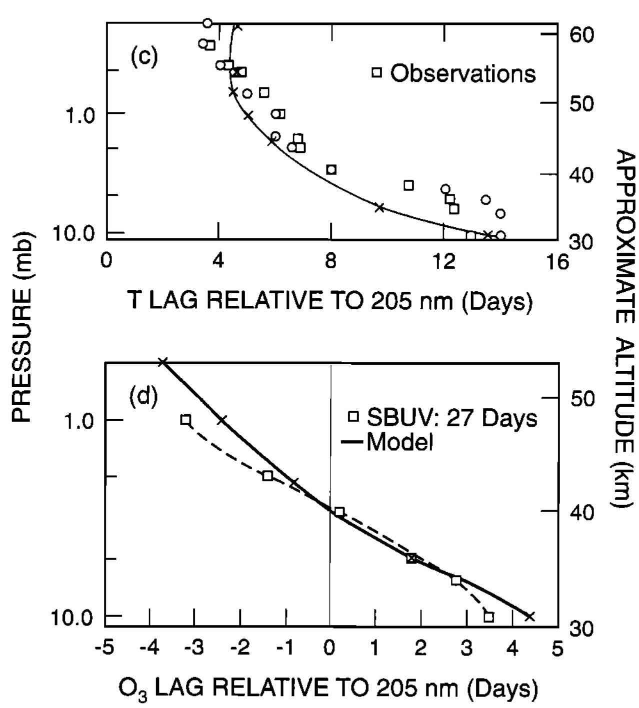


Fig. 5. (*a*) Ozone and temperature sensitivities relative to a 1% variation (27 days) in solar flux at 205 nm (expressed in % O_3/% UV flux, and in Kelvin/% UV flux, respectively). Calculated values representative of equatorial regions are compared as a function of altitude with LIMS, SBUV, and SAMS data analyzed by *Keating et al.* [1987] for the ± 20° latitude band. (*b*) Ozone sensitivity in the mesosphere relative to a 1% variation (27 days) in solar flux at Lyman-α (121.6 nm; expressed in % O_3/% Lyman-α). Calculated values are compared to the analysis by *Keating et al.* [1987], based on the SME data. (*c*) Calculated time lag (days) of the temperature signal relative to the 27-day solar forcing, compared with observations data [*Hood*, 1986; *Keating et al.*, 1987]. (*d*) Calculated time lag (days) of ozone signal relative to the 27-day solar forcing, compared with observational data (SBUV) analyzed by *Keating et al.* [1987]. Solid curves represent the model values, dashed curves the values deduced from observations.

could be due to differences in the adopted radiative codes or to the fact that our 2-D model simulates more accurately than earlier models some of the feedbacks between radiation and zonal mean dynamics.

Finally, the variability predicted for the ozone column abundance is presented in Figure 6. Assuming again that the solar variation at 205 nm is 3.3%, the maximum response in the vertically integrated ozone concentration, which is located near the subsolar point, is 0.25% (maximum sensitivity of 0.076 percent by percent). Thus even during the periods of highest solar activity when the 27-day solar signal is largest, the variation in the ozone column abundance in response to the 27-day solar variability should be less than 0.5%. The response of the ozone column lags the solar forcing by 2.5 days. The overall pattern does not differ significantly with season in response to changes in the intensity of meridional transport; the maximum ozone response follow, however, the seasonal variation of the subsolar point.

4. Middle Atmosphere Response to 11-Year Solar Cycle

Because the 11-year period is long compared to the time needed for the chemical system to reach a quasi steady state, the response of the atmosphere over a solar cycle can be derived by comparing two model simulations performed at steady state for solar minimum and solar maximum conditions, respectively. As indicated earlier, the variation in the solar actinic flux between these two model calculations is shown in Table 1. Figures 7*a* and 7*b* represent as a function of latitude and season the calculated change in the ozone column abundance for an increase in the solar flux from minimum to maximum conditions. In Figure 7*a* the lower solar variability of 3% between 208 and 265 nm is used, while in Figure 7*b* the higher value of 5% is adopted. In both cases, the change in the ozone abundance, which is of the order of 1.1–1.3% at the equator, increases with latitude, especially in winter and early spring.

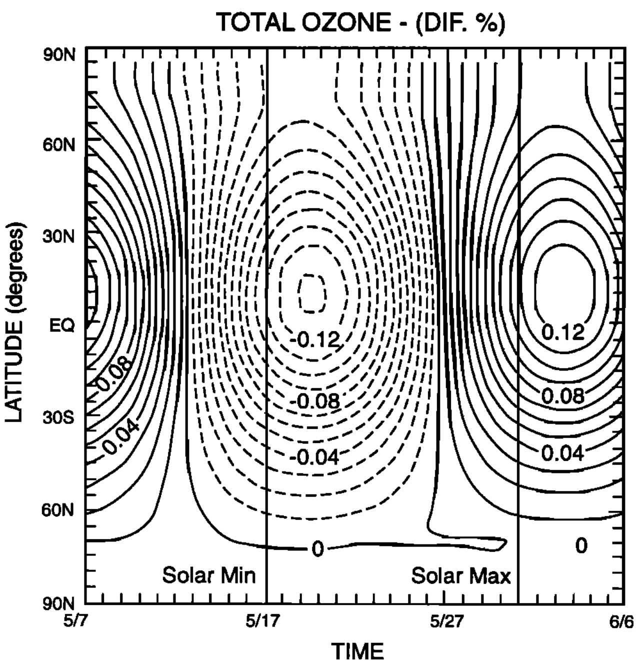


Fig. 6. Same as Figure 4, but for the response of the ozone column as a function of latitude.

This is a direct consequence of the strong downward circulation associated with planetary wave breaking during this period of the year, which transports to relatively low levels and high latitudes ozone-enhanced air masses from higher altitudes and lower latitudes. The difference in the two results is noticeable and puts constraints on the possible impact on total ozone of uncertainties in the solar variability in the spectral region of the ozone Hartley bands. Note that the difference between the results shown in Figures 7*a* and 7*b* is much smaller than the change in the forcing between 208 and 265 nm (from 3 to 5%), because a significant fraction of stratospheric ozone is produced by radiation at wavelengths shorter than 208 nm. Uncertainties in the flux variations over the spectral region of the O_2 Schumann-Runge bands ($\lambda < 200$ nm) affect mostly the layers in and above the upper stratosphere and hence also have a limited effect on the total ozone column abundance. The variability in the Herzberg continuum, especially in the atmospheric window between 190 and 210 nm has the largest impact on total ozone, and the long-term solar variability in this wavelength region seems to be fairly well established.

The variation from solar minimum to solar maximum conditions of temperature, ozone and other chemical constituents is shown in Figures 8*a* and 8*b* to 11*a* and 11*b* as a function of altitude and season at the equator and at 60°N,

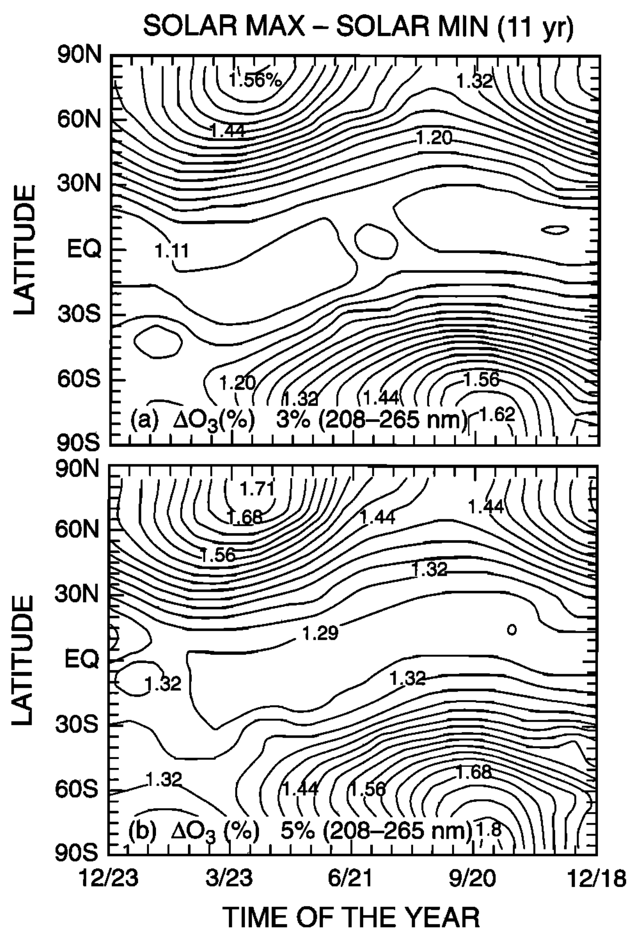


Fig. 7. Response as a function of latitude and time of the year (percent) of the ozone column abundance to the 11-year solar cycle. (*a*) Assumed variation in the solar flux between 208 and 265 nm, 3%; (*b*) same as Figure 7*a* but with a variation of 5% for the solar flux between 208 and 265 nm.

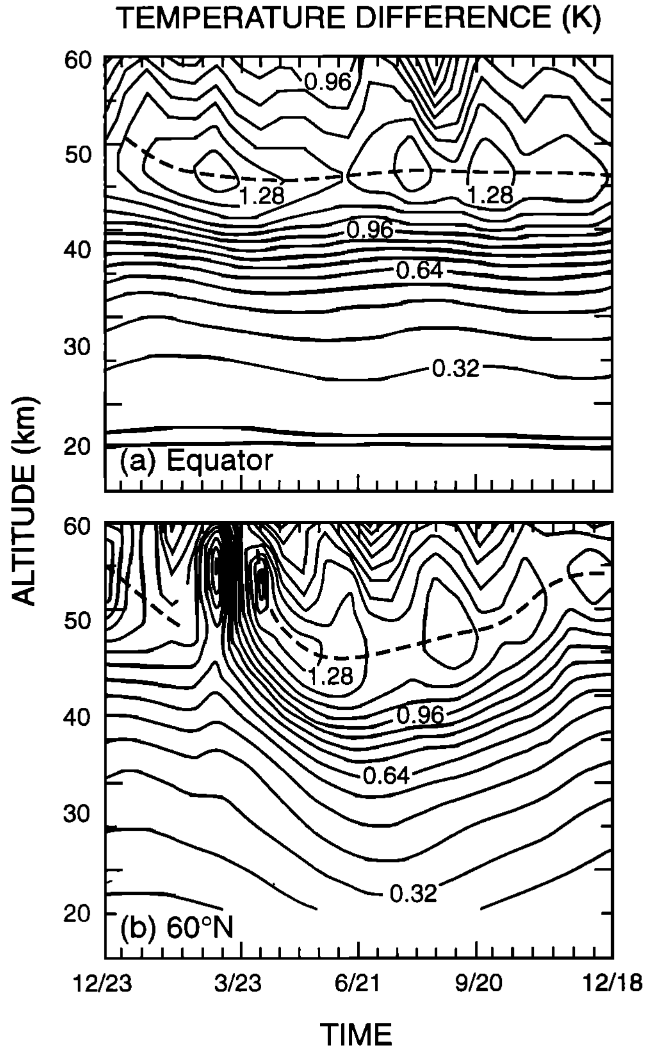


Fig. 8. Response to the 11-year solar cycle of temperature (Kelvin) as a function of altitude and time of the year: (*a*) at the equator, and (*b*) at 60°N.

respectively. In the case of the temperature (Figures 8*a* and 8*b*), the change over the solar cycle reaches approximately 1.4 K at the stratopause. It is relatively constant as a function of season in the tropics but varies significantly with season at 60°N. At this high latitude, the height of the maximum response decreases as solar radiation penetrates deeper into the atmosphere. Thus in the stratosphere at a given altitude, the temperature response is largest in summer. The variability in the zonal wind (not shown) calculated by the model is generally lower than 1 m/s and hence is not expected to substantially affect the vertical propagation of planetary and gravity waves. The eddy diffusion coefficients, which are parameterized in the model as a function of the mean zonal wind, are occasionally modified in a significant way, in response to changes in the vertical propagation and breaking of gravity waves. Such an effect is, for example, noticeable in the mesosphere at 60°N during March (see Figure 8*b*). Only simulations with high-resolution general circulation models will provide a more definite answer concerning potential dynamical perturbations.

The response of the ozone concentration to the 11-year cycle (Figures 9*a* and 9*b*) reaches a maximum of approximately 2–2.5% at 35–40 km altitude. Again, a strong seasonal variation in the response is seen at high latitudes, whereas it is entirely lacking at the equator. Moreover, the amplitude of the response decreases with altitude in the upper stratosphere and even becomes negative in the mesosphere. This feature in the model is a manifestation of the important temperature/ozone feedback in this region and of the role played by hydrogen compounds in the ozone budget. It is, however, different from the 11-year ozone variations deduced from the SBUV observations (Hood et al., submitted manuscript) which suggest that the maximum ozone response takes place near the stratopause. This discrepancy between models and observations, which needs to be further investigated, is perhaps explained by the short time period over which data have been collected (poor statistics) but could also be related to the inability of the models to accurately reproduce the ozone concentration at and above 40 km altitude [*Eluszkiewicz and Allen*, 1993]. Under this hypothesis, the "missing" process

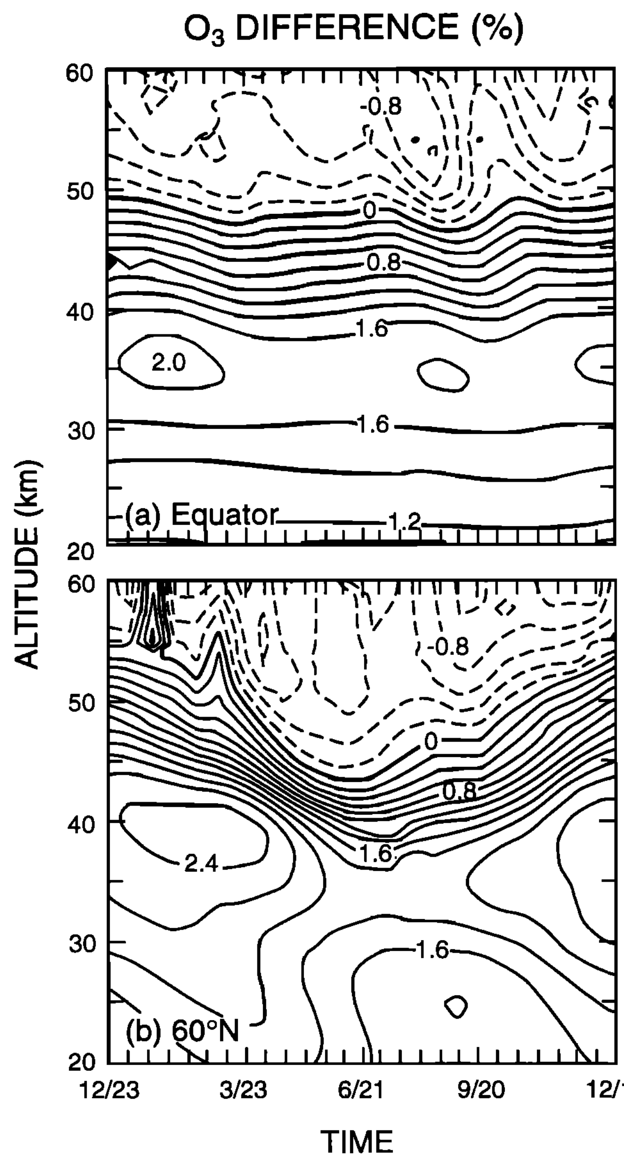


Fig. 9. Response to the 11-year solar cycle of the ozone concentration (%) as a function of altitude and time of the year. (*a*) at the equator; (*b*) at 60°N.

which would reconcile calculated and measured ozone concentrations should involve solar radiation at relatively short wavelengths (less than 200 nm). For example, the possibility that there is an atmospheric penetration of solar radiation in the spectral range of the O_2 Schumann-Runge bands that is stronger than commonly accepted should be investigated. Note, however, that the vertical distributions of the SBUV weighting functions are relatively broad (coarse vertical resolution), so that the reported ozone changes might not be very accurate. The fact that ozone variations are largest near the stratopause needs to be confirmed by the analysis of future data.

Hydroxyl radicals in the stratosphere are produced mainly by the reaction of water vapor with the electronically excited oxygen atom $O(^1D)$. Because water vapor is not significantly affected by solar ultraviolet radiation below 60 km, the response of the OH radical in the stratosphere is related mainly to that of the $O(^1D)$ atom. This latter atom is formed by the photolysis of ozone at wavelengths less than 310 nm. Above approximately 30 km, the concentration of $O(^1D)$ increases (by 3.2% at 45 km) from solar minimum to solar maximum conditions, but below this altitude, the response of $O(^1D)$ becomes negative, since the solar flux at these levels is reduced as a result of increased solar absorption by the enhanced concentrations of ozone in the upper stratosphere. An increase in the OH concentration of 1.6% and a decrease of approximately the same magnitude are calculated at 50 km and 20 km altitude, respectively (see Figures 10*a* and 10*b*). A 1.6% reduction in the OH density is also predicted at 60°N near 40 km during winter.

Finally, the response of the nitrogen family depends on the variations in the odd nitrogen (NO_y) source which is proportional to the densities of N_2O and $O(^1D)$. Since nitrous oxide is photolyzed by solar ultraviolet near 200 nm, the concentration of this gas decreases with increasing solar activity. As indicated earlier, the concentration of the electronically excited oxygen atom decreases in the middle and lower stratosphere. Above 50 km, the net destruction rate of NO_y is proportional to the photodissociation rate of NO [*Brasseur and Nicolet*, 1973]. The photolysis of nitric oxide is due mainly to the predissociation in the delta bands [*Cieslik and Nicolet*, 1973] and thus involves the solar radiation near 180–190 nm. In our model, its value increases with solar activity by approximately 8%. As a result of these processes, the total amount of odd nitrogen decreases in most parts of the middle atmosphere as solar activity increases. This conclusion may, however, have to be revised because the analytic expression of *Nicolet* [1979] used in our study to approximate the photodissociation frequency of NO does not account for the attenuation of solar radiation by the lines (delta bands) of NO itself [see *Frederick et al.*, 1983]. As pointed out by *Minschwaner and Siskind* [1993], the increase in the NO abundance in the thermosphere with increasing solar activity may lead to a reduction in the photolysis rate of NO in the mesosphere and hence to an increase (rather than a decrease) in the concentration of odd nitrogen in the upper stratosphere and in the mesosphere.

The most abundant species belonging to the odd nitrogen family in the lower stratosphere is nitric acid. This compound is formed by the reaction between NO_2 and OH and is destroyed by photolysis and reaction with the OH radical. The fact that the HNO_3 photodissociation increases with solar activity and that the total amount of odd nitrogen species is reduced when the solar UV flux increases explains the calculated response of HNO_3 shown in Figures 11*a* and 11*b*. Reductions of approximately 2.5% at 25 km and 3.5% at 30 km are calculated between solar minimum and solar maximum conditions. The amplitude in the HNO_3 response becomes even larger at higher altitude, where the absolute concentration of this atmospheric constituent becomes extremely small. No data are yet available to validate these model results. It should be noted, however, that a 27-day signal was detected by *Keating et al.* [1986] in the HNO_3 concentrations recorded by the LIMS in 1978–1979.

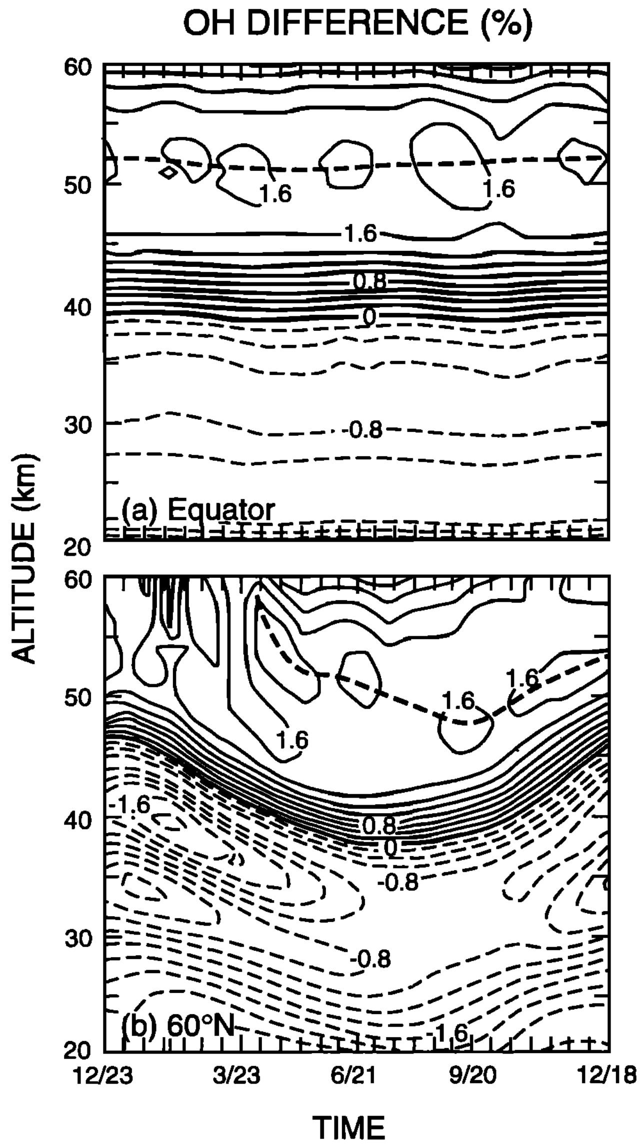


Fig. 10. Same as Figure 9, but for the hydroxyl radical.

A factor that could strongly influence the ozone response, and which has been discussed in detail by *Garcia et al.* [1984], *Brasseur* [1989], and more recently by *Huang and Brasseur* [1993] is the influence of solar-induced changes in thermospheric nitrogen oxides and their downward transport in the polar regions during winter. To analyze this possible influence, we have arbitrarily enhanced the concentration of NO_y at the upper boundary of the model (85 km) and have observed the response of lower atmospheric layers to this

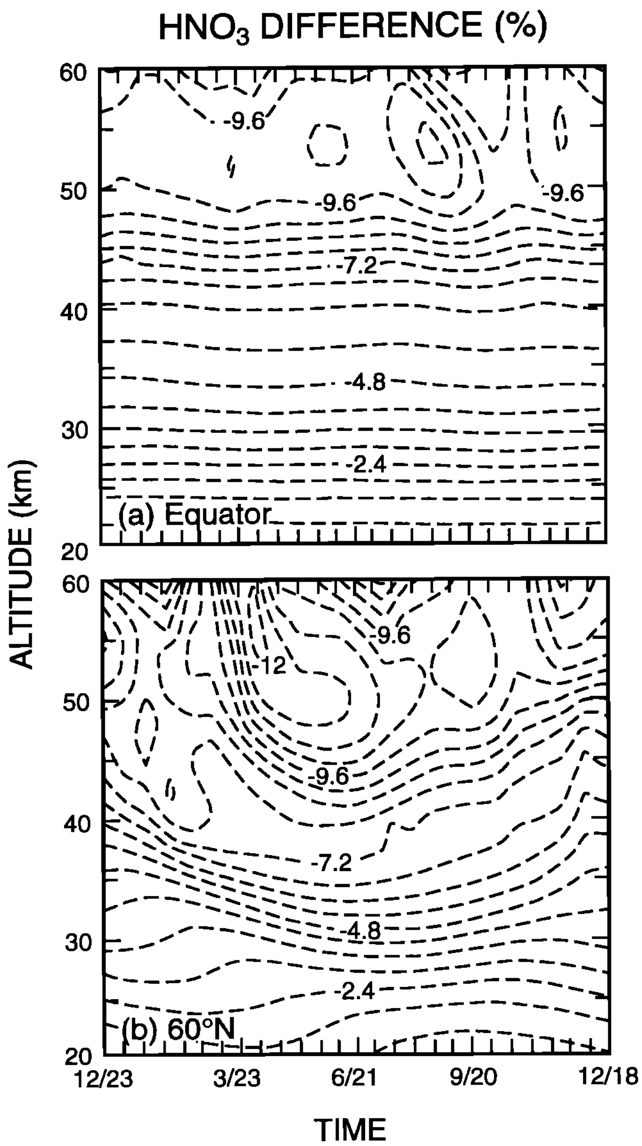


Fig. 11. Same as Figure 9, but for nitric acid.

perturbation. Figure 12*a* shows the relative change in the NO_y concentration at the level of the stratopause as a function of latitude and season when the thermospheric density of NO_y is increased uniformly by a factor 5 during all seasons. While the downward transport of thermospheric NO_y is inhibited at latitudes where the photolysis of NO and the recombination of N and NO lead to a rapid destruction of odd nitrogen, it influences the upper stratosphere at high latitudes in winter and early spring. Figure 12*b* represents the relative increase in the NO_y concentration as a function of height and time of the year at the particular latitude of 75°N. Although this calculation was performed for sensitivity purposes with a NO_y forcing which probably exceeds by a factor of 2 the observed one [*Roble and Emery*, 1983], it qualitatively shows the spatial and temporal extent of the thermospheric perturbation. The response of the ozone column abundance to the solar cycle (with the thermospheric intrusion of NO_y increased in this case by a factor 4) is shown in Figure 13. The amplitude of the variation over an 11-year period is not substantially changed in the tropics, compared to the case shown in Figure 7*b* where the effect of NO_y is ignored. At high latitudes, however, the change in the ozone column is somewhat reduced especially in the Antarctic region during spring (1.3% rather than 1.80%). It should be emphasized at this point that the model does not account for the heterogeneous processes which play a key role in the formation of the Antarctic ozone hole and perhaps in the observed ozone depletion at mid-latitudes. Model simulations performed when heterogeneous chemistry are taken into account produce very similar responses of the atmosphere to solar variability. The present simulations suggest that, even with large intrusions of nitrogen oxides from the thermosphere, only limited destruction of polar ozone takes place in the lower stratosphere.

5. DISCUSSION AND CONCLUSIONS

The calculated response of stratospheric ozone and temperature to the 27-day solar forcing (sensitivity to 205-nm

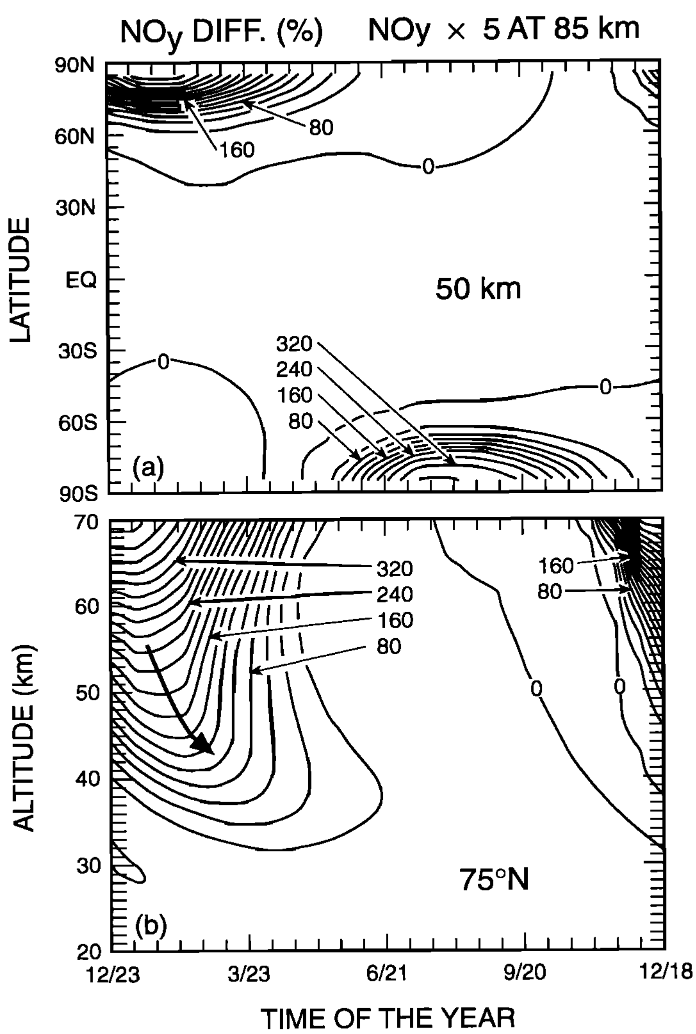


Fig. 12. Intrusion of nitrogen oxides from the lower thermosphere into the stratosphere. Relative variation (percent) in the concentration of odd nitrogen (NO_y = NO + NO_2 + NO_3 + 2 N_2O_5 + HNO_3) when the concentration is arbitrarily doubled at 85 km. (*a*) Percentage increase in NO_y at 50 km altitude as a function of latitude and season. (*b*) Percentage increase in NO_y at 75°N as a function of altitude and season.

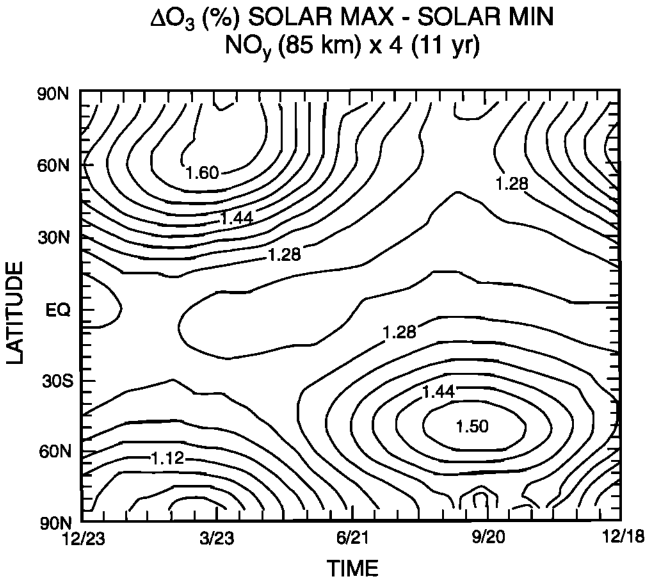


Fig. 13. Relative variation (percent) of the ozone column abundance to the 11-year solar cycle. In addition to the change in the solar flux from minimum to maximum conditions, the concentration of NO_y at 85 km is multiplied by 4 to account for an enhanced ionic production of nitric oxide in the lower thermosphere.

solar flux and phase lag) is in relatively good agreement with the corresponding values derived from satellite observations. In the mesosphere, the sensitivity of ozone to changes in solar radiation at Lyman-α is also well reproduced by the model. Because several radiative processes specific to the mesosphere are not treated in detail, the response of the temperature above 60 km has not been considered in the present study. However, as suggested by the work of *Huang and Brasseur* [1993], the anticorrelation in the temperature response (maximum amplitude) and the ozone response observed at 70 km [*Keating et al.*, 1987] cannot be explained by current atmospheric models. It remains difficult to understand from pure chemical and radiative considerations why a large positive temperature response is produced in a layer where a large negative ozone response is predicted and observed. Exothermic chemical heating or processes specific to nonlocal thermodynamic conditions (such as the quenching between CO_2 and atomic oxygen, which converts radiative energy into thermal energy in the mesosphere), which are affected by solar activity, do not explain the observed negative response in the temperature near 70 km. The problem needs to be studied.

In the case of long-term solar variations (11-year cycle), the agreement between models and observations is not satisfactory. The amount and quality of the data are insufficient to unambiguously derive the ozone and temperature responses. Available data suggest, however, substantial differences with model calculations. For example, the observed ozone response

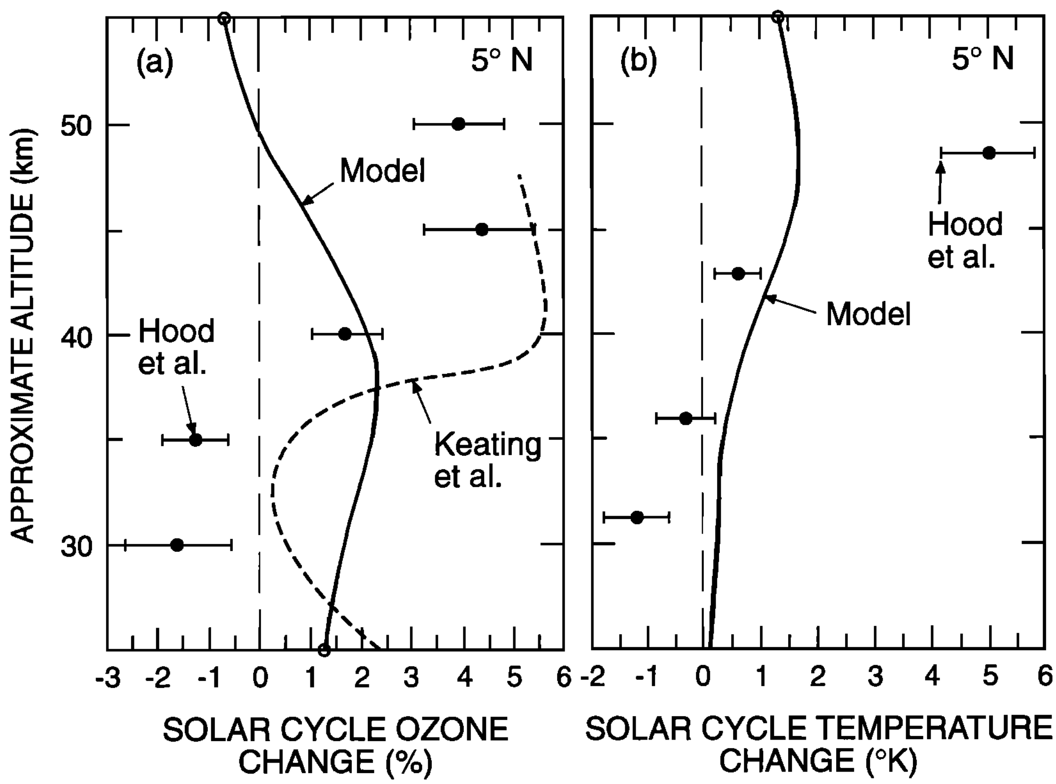


Fig. 14. Comparison between calculated and observed variations in (*a*) ozone concentration and (*b*) temperature at 5°N between 25 and 55 km altitude, in response to the 11-year solar cycle. The ozone variation calculated by *Keating et al.* [1993*a*] is based on SBUV and SAGE II data. Hood et al. (submitted manuscript) based their analysis on SBUV observations for ozone and NMC analyses for temperature.

near the stratopause, at least in the tropics, is a factor 2–3 higher than the calculated response, while the opposite is found near 30–35 km altitude [*Keating et al.*, 1993*a*; Hood et al., submitted manuscript]. The observed temperature response in the same region of the atmosphere also seems to be different than predicted by models (Hood et al., submitted manuscript). Not only do the absolute values of the ozone and temperature responses not match, but the variation with height of these responses (shown in Figures 14*a* and 14*b*) is very different when calculated in models or retrieved from observations. It should be remembered, however, that SBUV ozone and NMC temperature information is relatively poor, especially above 30 km, where the weighting functions of the corresponding instruments are broad. The comparison between models and data may not be favorable simply because the data themselves have some uncertainties due to insufficient vertical resolution. The problem needs, however, to be investigated in future studies. An explanation for the established discrepancies could involve chemical, radiative, as well as dynamical processes which are not yet fully understood. The role of dynamics in the atmospheric response to solar variability should be investigated with the aid of three-dimensional general circulation models in which a detailed formulation of coupled chemical and radiative processes is implemented.

Finally, it should be stressed that the response of the atmosphere to solar variability is very different depending on the period of the solar forcing, so that calculated or observed short-term effects cannot be directly used to derive long-term effects. For example, our model calculations suggest that, for identical amplitudes in the solar forcing at all wavelengths, the amplitude of the change in the ozone column abundance for an 11-year forcing is approximately 2.5 and 3.5 times larger at the equator and at mid-latitudes, respectively (equinox), than that for a 27-day forcing (see also *Keating et al.* [1993*b*].

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