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Observed Ozone Response to Variations in Solar Ultraviolet Radiation

Abstract. *During the winter of 1979, the solar ultraviolet irradiance varied with a period of 13.5 days and an amplitude of 1 percent. The zonal mean ozone values in the tropics varied with the solar irradiance, with an amplitude of 0.25 to 0.60 percent. This observation agrees with earlier calculations, although the response may be overestimated. These results imply changes in ozone at an altitude of 48 kilometers of up to 12 percent over an 11-year solar cycle. Interpretation of ozone changes in the upper stratosphere will require measurements of solar ultraviolet radiation at wavelengths near 200 nanometers.*

The composition of the stratosphere, including the concentration of ozone (O_3), is controlled by photochemical processes that are driven by solar radiation (*1*). Humphreys suggested (*2*) that variations in the solar output could modify the O_3 content of the stratosphere. Numerous photochemical calculations, including those carried out recently (*3, 4*), have confirmed this prediction. Many investigators (*5*) have used observational data to search for relationships between solar parameters and amounts of O_3, but the results have been controversial and not always convincing, usually because the measurements have been relatively limited in number, noisy, or subject to long-term drift or to interference by other geophysical phenomena. We report here the clearest and most quantitative relationship between solar ultraviolet radiation and stratospheric O_3 thus far.

The O_3 response to solar radiation results from the reaction

$$O_2 + h\nu \rightarrow O + O$$

at a wavelength $\lambda \leq 242$ nm, which is followed by the reaction

$$O + O_2 + M \rightarrow O_3 + M$$

The destruction of odd oxygen ($O + O_3$) is less sensitive to ultraviolet changes (*1*). Thus, we expect an increase in solar ultraviolet radiation for $\lambda < 242$ nm to cause an increase in O_3.

The Limb Infrared Monitor of the Stratosphere (LIMS) experiment operated on the Nimbus 7 spacecraft from 25 October 1978 to 28 May 1979. It measured the infrared emitted by the earth's atmosphere at the limb, from which vertical distributions of temperature, O_3, and three other trace gases were derived (*6*). The individual O_3 profiles appear to be accurate to 10 percent and—more important here—precise to ≤ 5 percent (*7*). The Solar Backscatter Ultraviolet Experiment (SBUV) also flew on Nimbus 7, measuring the solar spectrum and its variation (*8*).

The LIMS profiles were derived every 4° of latitude, from 64°S to 84°N. We then analyzed the data at these latitudes into daily values for the zonal mean, plus the amplitudes and phases of the first six harmonics around the latitude circle, using a Kalman filter method (*9*). We have looked for solar effects on the zonal mean O_3 values.

It has long been recognized that rotation of active solar regions should lead to a modulation of ultraviolet solar irradiance. Measurements by Heath and his co-workers (*10*) have shown that the relative amplitude of this modulation is a stable function of λ, and so changes at one λ can be related to those at another. We have used SBUV measurements at 205 nm as an indicator of the variations between 190 and 220 nm (*3*) to which atmospheric O_3 is most sensitive. Comparison could be made to the sum of variations at all wavelengths, weighted by their calculated photochemical efficiency, but this would introduce model dependence which we wished to avoid.

Beginning in early 1979, the solar (S) measurements show several oscillations with a period of ~13.5 days, due to two groups of strong solar-active regions about 180° solar longitude apart and the sun's 27-day rotation (*10*). These data are shown in the lower panel of Fig. 1a.

The LIMS zonally averaged O_3 for the equator at 5 mbar (~36 km altitude) is also shown as a function of time in the upper panel of Fig. 1a. The largest variations are due to seasonal effects, which obscure ultraviolet-related effects.

Since the sun provides an oscillating signal of narrow bandwidth, we used statistical spectral methods to look for an O_3 response in the same frequency range. A power-spectral analysis of the S data (one point per day for the 139 days from 10 January to 28 May) shows a large peak at a frequency near 0.074 day^{-1}, or a period of 13.5 days, as expected. Similar analysis of the O_3 data for a range of altitudes and latitudes shows the power predominantly at low frequencies but with a distinct peak in the frequency range 0.06 to 0.075 day^{-1}.

Cross spectra between the two series were calculated (*11*) and smoothed with a Hamming window five modes wide. These spectra invariably showed significant peaks in the squared coherence (C^2) at the 13.5-day period. Almost all $C^2 > 0.5$ occurred at and above the altitude of the 10-mbar levels, with the largest single region between 24°S and 24°N. This is not surprising, since traveling waves (*12*) in mid-latitudes, which are unrelated to solar effects, would be expected to have considerable power of the same frequencies. The subsequent discussion focuses on this tropical region. Because atmospheric variability is a source of noise whereas the true O_3 response to the ultraviolet variation should be quite similar at all tropical latitudes, we averaged the C^2 spectra for each pressure level. The values for 5 mbar are shown by the solid line in Fig. 1b. Cross spectra were also calculated between S and a white-noise sequence. The range of values is shown by the dashed lines in Fig. 1b. The peak at 13.5 days is 0.68, several standard deviations above a random effect. Almost all other frequencies are within or close to the range of random effects. These significant peaks at the several pressure levels indicate a strong connection between ultraviolet variation and O_3 response.

The phase lag ϕ between the 13.5-day O_3 and S variations was also estimated. It is less than 1 day above 40 km, whereas below that level O_3 lags S by intervals that increase with increasing pressure. Theoretically, ϕ (in days) = $13.5/2\pi \tan^{-1} (2\pi\tau/13.5)$, where the response time to the perturbation τ is half the conventional O_3 lifetime. The uncertainty of the phase determination encompasses the theoretically predicted lags (for example, 2.1 to 2.5 days at 10 mbar), although the observed lags are systematically shorter (*13*). Below 10 mbar there are too few latitudes with high C^2 to give meaningful results.

To determine the ratio of O_3 response to ultraviolet change, the cross-spectral gain must be calculated. However, the Hamming window acting on the red O_3 spectrum will overestimate the O_3 power at the frequency of interest. To reduce this effect, both data sets were digitally filtered with a band-pass filter centered at 0.075 day^{-1}, having half-power points at 0.050 and 0.100 day^{-1} (20- and 10-day periods, respectively); linear trends were removed, and the variation was expressed as a percentage of the mean. The variations in the resulting S and O_3 time series are small, ± 1 percent or less, but the peaks line up very closely in time, again showing the O_3 response to the solar ultraviolet variation.

Cross spectra of the filtered S and O_3 time series (90 points), smoothed with a five-mode Hamming window, were calculated. For latitudes and pressures at which C^2 at a frequency of 0.077 day^{-1} was ≥ 0.6, we calculated the gain, which

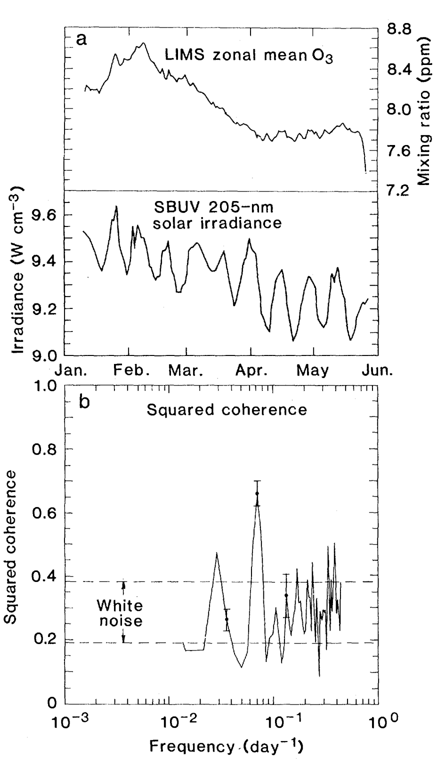

Fig. 1. (a) Zonal mean O_3 at 5 mbar (~35 km altitude) at the equator measured by LIMS and solar ultraviolet flux at 205 nm, measured by SBUV, from January to May 1979. (b) Squared coherence between LIMS O_3 and solar ultraviolet data, averaged over tropical latitudes.

is the ratio of the percentage change of O_3 to the percentage change of 205-nm solar ultraviolet irradiance. The average gains are presented in Fig. 2, with error bars indicating one standard deviation of the individual determinations. The line is dashed where only six latitudes show $C^2 > 0.6$. The gains show a very clear and significant response which increases from ~0.17 at 30 km in the low stratosphere to ~0.38 at 54 km in the low mesosphere, with larger values above. To our knowledge, these results are the clearest and most quantitative demonstration to date of the effects of solar variation on stratospheric O_3.

We are unaware of any calculations of the O_3 response to 13.5-day ultraviolet variations based on the use of Heath's observed spectra (*10*). However, Lean's (*14*) models indicate that the shape of the relative spectral variation of the 11-year solar cycle is very similar to that of these 13.5-day variations, although the amplitude is larger. Thus, to a first approximation, we can compare our results with solar cycle calculations of Brasseur and Simon (*3*), scaled to our smaller ultraviolet variations; these calculations are labeled "model" in Fig. 2.

There is a significant difference above 0.7 mbar, where this model predicts a decreasing response with altitude while the observed response increases. The observed values are insensitive to details of the data treatment but close to the detection limit. The theoretical decrease stems from several factors, including the effects of solar increases at $\lambda > 250$ nm, which accelerate O_3 destruction and warm the atmosphere, thereby reducing O_3 through the temperature dependence of reaction rates (*15*). However, the wavelength dependence observed by Heath and his co-workers (*10*) shows no appreciable continuum variation for $\lambda \geq 260$ nm. This would reduce the predicted decrease of sensitivity above 1 mbar. Consistently, LIMS detected no temperature change ≥ 0.15 K.

Below 0.7 mbar the agreement between the observations and the calculated values is surprisingly (and fortuitously) good, in view of the simplified wavelength dependence used by Brasseur and Simon (*3*), the (weak) dependence of the LIMS power spectrum on the details of the Kalman filter, and slight amplitude reduction as ϕ increases. Garcia *et al.* (*4*), using S variations close to those observed, predict a maximum O_3 sensitivity of 0.22 at 10 mbar with decreasing values above. Thus, presently understood photochemistry may not be completely consistent with our results, in magnitude or even in altitude variation. Some of the differences, of course, may be due to differences in the important dynamical and radiative processes for 13-day and 11-year time scales.

The present results may also be used to predict the O_3 variation over the solar cycle, if there is an estimate of the change at 205 nm. In Lean's work (*14*), the model solar cycle minimum-to-maximum change for this wavelength is 27 percent, which suggests a variation of ~12 percent at 1 mbar, slightly larger than indicated (*3*) in the tropics. This argument also leads to variations of ~8 percent for the layer from 32 to 46 km. This is reasonably consistent with Umkehr data (*16*), which, after correction for the effects of stratospheric dust from Mount Agung (*17*), indicate changes of 0 to 9 percent for this layer. The LIMS results would also imply a variation in total column O_3 slightly smaller than the 3.5 percent predicted by Brasseur and Simon (*3*), again generally in the range shown by Angell and Korshover (*16*).

Thus there appears to be a rough consistency between the present results, the photochemical model of Brasseur and Simon (*3*), Lean's model of 11-year solar variability (*14*), and corrected Umkehr observations of O_3 change. An immediate conclusion is that any attempt to measure the effects on upper stratospheric O_3 resulting from human activities, such as the release of chlorofluoromethanes (*1*), will need corrections for solar-induced changes. This will require long-term observations of the solar spectrum or, at a minimum, measurements at a few well-chosen wavelengths.

Finally, there has been a recent increase in interest in the effect of the variation of the solar output on the

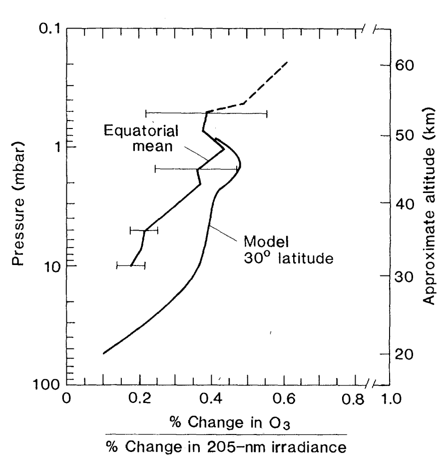

Fig. 2. Ratio of the percentage O_3 change to the percentage solar irradiance change at 205 nm obtained from this study compared to the model results of Brasseur and Simon (*3*). The dashed line indicates values close to the detection threshold.

earth's atmosphere (*18*). Clear connections are seen in the upper atmosphere (above 90 km), but below that it has been difficult to establish relationships based on the use of earlier measurements. The results presented here clearly demonstrate such a solar-terrestrial connection in the stratosphere.

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Species Formation Through Punctuated Gradualism in Planktonic Foraminifera

Abstract. *Detailed analysis of evolutionary changes in a 10-million-year long Late Neogene lineage of planktonic foraminifera has revealed a pattern that is not consistent with either the gradualistic or the punctuational model of evolution. The lineage was in stasis over a considerable part of its total duration but underwent relatively rapid, but not geologically instantaneous, gradual morphologic change that did not lead to lineage splitting. The term* punctuated gradualism *is suggested for this evolutionary modality.*

The question of whether new species arise through the slow and gradual phyletic transformation of entire populations leading to an unbroken, gradational chain of species (orthogenesis or phyletic gradualism) or whether they evolve rapidly (on a geologic time scale) in genetically isolated subpopulations (allopatric speciation or punctuated equilibrium) (*1*) has been discussed for more than a decade. It remains unclear which, if any, of these models predominates in the paleontological record. Many factors contribute to this uncertainty, but the most important is the scarcity of detailed quantitative studies of phyletic changes through long intervals of time in most fossil groups. Rather than forcing poorly documented patterns of evolutionary change to fit theoretical models, the evolutionary models should be formulated from studies of existing fossil lineages.

Evolutionary studies of fossil lineages, properly conducted, are laborious, but the time spent in generating the necessary data base should pay off by increasing our understanding of the nature of evolutionary patterns as well as of possible differences in patterns among organism groups. Morphologic changes need to be quantified from analyses of large numbers of specimens in long continuous sequences spanning several million years. Stratigraphic completeness and time resolution for a particular sampling density need to be evaluated (*2*). Patterns need to be tested statistically, for example, for precision of data, unidirectionality, stasis, and stepwise change. The geographic validity of an observed pattern must be confirmed to rule out the possibility that it has a nongenetic origin (migration).

We have analyzed the warmwater planktonic foraminiferal lineage *Globorotalia tumida* through the last 10 million years of its evolutionary history (Late Miocene–Recent) in DSDP Site 214 from the southern Indian Ocean (11°S, 89°E; water depth 1665 m). We were interested in determining long-term patterns of evolutionary change and, particularly, the evolutionary transition from *G. plesiotumida* to *G. tumida* across the Miocene-Pliocene boundary. We examined 105 samples taken at 10- to 30-cm intervals across the Miocene-Pliocene boundary and at about 2-m intervals in other parts. Sampling resolution was very good, between 5×10^3 and 15×10^3 years across the boundary and 2×10^5 years otherwise. Biochronologic analyses indicate a complete sequence across the Miocene-Pliocene boundary, whereas about 3 million years of sediments appear to be missing in the Late Miocene (in the interval between 10.4 and 6.4 million years ago).

About 50 specimens were picked at random from each sample. The specimens were oriented in an edge view. This view permits measuring of the elongation and inflation of the test; both are known to be responsible for evolutionary change in this lineage (*3*). The outline contour of the shape in this orientation was analyzed with the use of eigenshape analysis (*4*). This method represents the greatest proportion of variation observed among a collection of shapes by the least number of different shapes. The shape variable in eigenshape analysis is independent of size, which eases comparisons of different sized specimens.

The Late Miocene part of the sequence (*G. plesiotumida*) displays some variation in mean morphotype (Fig. 1), but most of this variation is clearly within the range of the precision of the data. Overall, these fluctuations were nondirectional and did not lead to any net phyletic change (*5*). Populations were thus more or less in stasis for about 5 million years. In the latest Miocene (about 5.6 million years ago) the shape began to change (at about the sample denoted 2 in Fig. 1); it changed gradually across the Miocene-Pliocene boundary, completing the transition from *G. plesiotumida* to *G. tumida* in the earliest Pliocene (about 5.0 million years ago; sample denoted 3). The rate of evolution was not constant during the transition (*6*); rather, the transition was marked by irregular fluctuations in morphotype in one general direction that caused the net change (*7*). This pattern appears to be a common mode of evolutionary change in those few lineages that are well documented through quantitative analysis (*6*).

Raup and Crick (*8*) have suggested the importance in evolutionary studies to