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Attribution of decadal variability in lower-stratospheric tropical ozone

Daniel R. Marsh[1] and Rolando R. Garcia[1]

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[1] The variability of ozone in the lower stratosphere in a climate-chemistry model is investigated by multiple regression analysis. The model includes forcing due to changes in solar irradiance and to anomalies in sea surface temperatures (SST). When ozone calculated for the period 1979–2003 is regressed against time and 10.7 cm radio flux (f10.7), the regression coefficient of f10.7 at 52 hPa is 2.8% per 100 flux units. This decreases to 1.8% if a lagged index of El Niño–Southern Oscillation (ENSO) is included in the regression model, and to 0.78% if the period of analysis is extended to 1950–2003. The last value is in good agreement with simulations of fixed solar maximum vs. solar minimum conditions that do not include SST variability. These results suggest that some of the decadal variability in tropical ozone previously attributed to solar variability may instead be related to the occurrence of ENSO events.

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1. Introduction

[2] Multiple linear regression is often used to extract the solar signal from observed ozone time series [see, e.g., *McCormack and Hood*, 1996; *Wang et al.*, 1996; *Randel and Wu*, 2007; *Soukharev and Hood*, 2006]. In the study by *Soukharev and Hood* [2006] (hereinafter referred to at SH06), regression analysis of observations from three satellite data sets was used to extract the response of stratospheric ozone to the solar cycle. They found a statistically significant increase in the tropical lower stratosphere of 2–4% between solar minimum and solar maximum. SH06 contrasted this response to four model predictions that showed much smaller solar signals, in the range of 0.5 to 1.5%. While the models chosen by SH06 range from older two-dimensional models to newly developed, fully interactive climate-chemistry models (CCM), their response to solar forcing is remarkably similar. A recent study comparing changes in ozone for fixed solar forcing using version 3 of the Whole Atmosphere Community Climate Model (WACCM3) by *Marsh et al.* [2007] also failed to show a large ozone response in the lower stratosphere, consistent with the earlier model results reported by SH06. This raises the question, why do these simulations not show a large ozone response to solar variability in the lower stratosphere? SH06 suggested that the mechanism of the enhanced response is dynamical in nature, and could be related (indirectly) to solar induced changes in the upper stratosphere. While such a possibility cannot be ruled out, it is puzzling that CCMs such as WACCM3 that include a stratosphere and mesosphere do not reproduce such behavior.

[3] Terms used in regression analysis can include a linear trend in time as well as proxies for the quasi-biennial oscillation (QBO), volcanic aerosols and, of course, solar irradiance. One of the assumptions of this approach is that the fitting indices are linearly independent. If this is not the case, then erroneous attribution of the driving mechanisms may occur. Figure 1 shows the cross-correlation coefficient between the monthly mean 10.7 cm radio flux (f10.7) and the Niño 3.4 index (N3.4; the standardized mean sea surface temperature in the region 5°S–5°N and 120°W–170°W) as a function of lag for the period 1979–2003. The error bars indicate the 2σ uncertainty of the correlation coefficients, and have been determined using block bootstrap resampling [see, e.g., *Wilks*, 2006], with a block size of 12 (to allow for significant autocovariance on time scales up to one year). The cross-correlation is nearly significant at the 2σ level for lags near -6 months, and has values approaching -0.2 over the range of lags -5 to -9 months. On the other hand, the cross-correlation between f10.7 and N3.4 over the longer period 1950–2003 (not shown) is nearly zero at all negative time lags.

[4] The existence of a significant cross-correlation between f10.7 and N3.4 in 1979–2003 implies that ozone variability due to the latter could be erroneously ascribed to the former when this period is analyzed, especially if the response of ozone to N3.4 in the lower stratosphere occurs with a lag of a few months (since this will tend to bring the ozone signal into phase with f10.7). We show in what follows that this is precisely what happens in WACCM3 and that, if one includes a lagged ENSO index in the multiple regression analysis, the apparent response of ozone in the lower stratosphere to solar variability is reduced sharply. This may explain much of the discrepancy between regression analyses of model output and those based on data when interannual variability in tropical sea-surface temperature (SST) is not taken into account.

2. Simulations

[5] WACCM3 is a fully-interactive CCM with a vertical domain that extends from the surface to the lower thermosphere (~140 km). The response of the model to changes in irradiance over the solar cycle using fixed solar maximum vs. solar minimum conditions is discussed by *Marsh et al.* [2007]. In this study we perform multiple linear regression on output from transient simulations of the period 1950–2003. The experimental setup is described in detail by *Garcia et al.* [2007]. Briefly, a set of four simulations at horizontal resolution of 4° latitude by 5° longitude were performed with realistic forcing, i.e., driven by variable spectral irradiance, SST, and changes in concentrations of key surface constituents. The simulations have been shown

[1]National Center for Atmospheric Research, Boulder, Colorado, USA.

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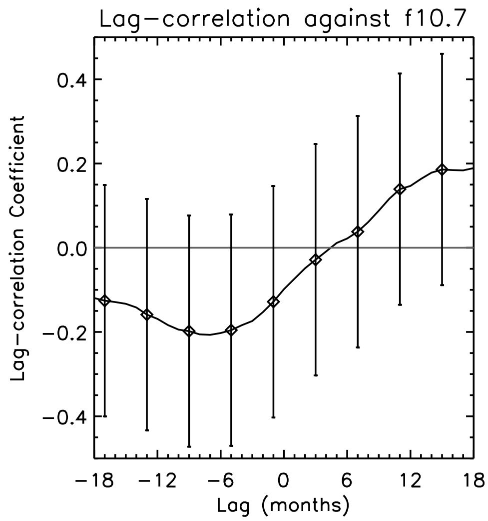

Figure 1. Lagged cross correlation of monthly-mean N3.4 and f10.7 indices over the period 1979 to 2003. Positive lag indicates that f10.7 leads N3.4. Error bars indicate the 2σ uncertainty of the correlation coefficient.

to be in reasonable agreement with the observational record [*Eyring et al.*, 2006; *Garcia et al.*, 2007]. For this analysis, model output was averaged by month, longitude, and latitude band, and over all ensemble members.

[6] Figure 2 shows the time-series of deseasonalized ozone at two model pressure levels for the Tropics (the latitude band ±24°); also shown are f10.7 and N3.4. Ozone at 4.3 hPa shows a long-term downward trend that is primarily the result of catalytic loss due to increasing chlorine levels in the stratosphere. In addition to the long-term trend, there is clear decadal variability that appears correlated with the solar cycle. At 52 hPa (~21 km), on the other hand, the solar signal is difficult to discern; however, at this level much of the variability is evidently related to ENSO, with relatively large (>5%) reductions in ozone mixing ratio following major warm events.

3. Regression Analysis

[7] To quantify the solar effect on ozone throughout the stratosphere, we use the standard technique of multiple linear regression. We fit the ozone deseasonalized time series to the following form:

$$O_3(t) = \overline{O}_3 + \alpha t + \beta f10.7(t) + \gamma N3.4(t-\tau), \qquad (1)$$

where t is time in months, $\overline{O}_3$ is the time-mean ozone, and τ is a lag chosen to maximize the projection of ozone on the ENSO signal, as explained below. The regression does not include a term for the QBO since WACCM3 does not generate one internally. This omission is actually an advantage for our analysis because it avoids possible aliasing of the solar and QBO signals (see *Lee and Smith* [2003] for a discussion of this topic).

[8] When regressing ozone time-series against proxies of possible drivers of ozone variability, consideration should be given to the time it takes for the effect to reach the region being analyzed. For example, in the study of *García Herrera et al.* [2006] it was shown that it takes several months for the ENSO temperature signal to propagate to the tropical lower stratosphere. To estimate the appropriate lag for the N3.4 index used in the multiple regression (1), cross-correlations of WACCM3 ozone and N3.4 were calculated; values for 52 hPa, together with uncertainty estimates obtained by block bootstrap resampling, are shown in Figure 3. Approximately 4–5 months following a warm ENSO event there is a significant negative response in ozone at 52 hPa; a similar lagged response is seen at all levels between 100 and 20 hPa (not shown). Nearly identical cross-correlations are calculated for temperature as for N3.4, and both are likely due to changes in the residual mean circulation [cf. *García Herrera et al.*, 2006]. This interpretation is supported by the positive response calculated for the residual-mean vertical velocity (also shown in Figure 3), which maximizes 2 to 3 months after a warm ENSO event. The anomalous upwelling induced by ENSO causes adiabatic cooling and simultaneously a reduction in ozone because the vertical gradient of ozone mixing ratio is strongly positive at 52 hPa. Further, the lag between the vertical velocity response and the responses in ozone and temperature is consistent with the long relaxation time scales of both odd oxygen ($O + O_3$) and temperature in the lower stratosphere, which are of the order of 2–3 months [*Brasseur and Solomon*, 1986; *Randel et al.*, 2002]. All of this suggests that the lag of the N3.4 index, τ in (1), should be set to a value between 4 and 5 months at 52 hPa. In the regressions presented here we set τ equal to the lag for which the cross-correlation between N3.4 and ozone is largest (typically between 3 and 6 months). Note that, if a time lag were not included in the regression onto N3.4, the projection of ozone on N3.4 would be much reduced, as implied by Figure 3 (and the projection on f10.7 consequently increased).

[9] Figure 4 shows amplitudes of the solar and ENSO regression coefficients for WACCM3 tropical ozone derived from the multiple linear regression (1); 2σ uncertainties obtained by block bootstrap resampling are indicated by shading. Over the period 1950–2003, the solar signal is

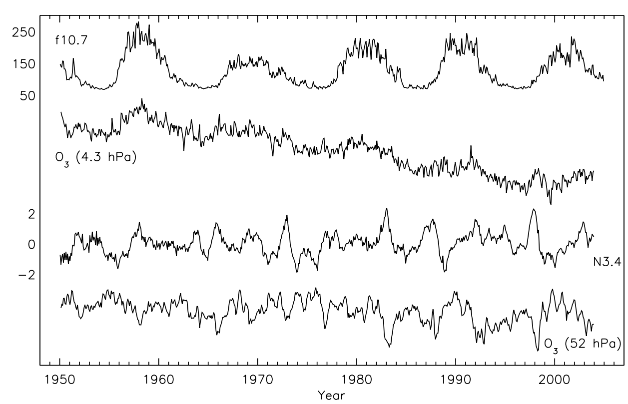

Figure 2. Monthly-mean time series of f10.7 (upper left axis); zonal-mean ozone (ppmv) at 2.57 hPa (upper right axis); N3.4 (lower left axis); and zonal-mean ozone (ppmv) at 51.7 hPa (lower right axis). Ozone values are tropical averages over ±24°.

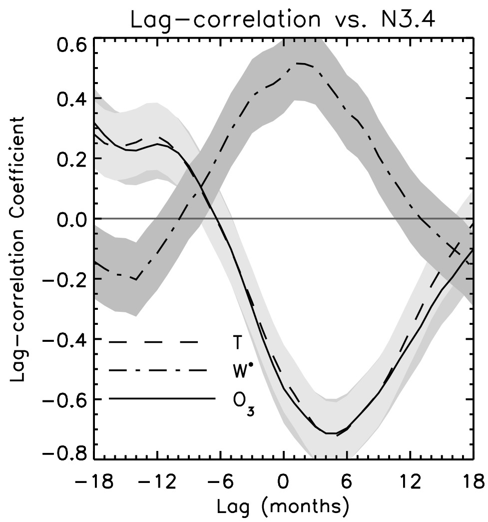

Figure 3. Cross correlation of 52 hPa ozone, temperature and residual-mean upwelling computed with WACCM3 versus N3.4. The WACCM3 variables are tropical averages over ±24° between 1950 and 2003. Positive lag indicates that N3.4 leads the model variables. Shaded areas indicate the 2σ uncertainty of the correlation coefficients.

0.74% per 100 units of f10.7 at 52 hPa when both f10.7 and N3.4 are used in the regression analysis, and remains essentially unchanged (0.78% per 100 units) when only f10.7 is used in the regression, indicating that the two proxies are uncorrelated over this period. Note, however, that the estimated uncertainty (denoted by the darker shaded region in Figure 4) increases in the lower stratosphere, where a sizable fraction of the ozone variance is not fit when N3.4 is excluded from the multiple regression. When the shorter period 1979–2003 is analyzed, the solar signal at 52 hPa is 1.8% per 100 f10.7 units if both proxies are used in the regression, but increases to 2.8% when only f10.7 is used. The last number is comparable to that obtained by SH06 for the same period using SBUV data (see SH06's Figure 8), while the first is comparable to that calculated in the models reviewed in SH06. These models contrast fixed solar maximum vs. minimum conditions, so that SST variability does not play a role in the results. (Note, by the way, that SH06 present their results as percentage change over the solar cycle, so our values must be multiplied by 133/100, where 133 is the mean range of f10.7 between solar minimum and maximum, when comparing to SH06). All of this suggests that, in the relatively short record 1979–2003, there is spurious projection of low-frequency variance onto f10.7 which should more plausibly be attributed to ENSO.

[10] The N3.4 regression coefficient itself is large and negative in the lower stratosphere over 1979–2003, with a value of −2.96% per unit of N3.4 at 52 hPa. This is within 10% of the value obtained for 1950–2003 (−2.66%) indicating that the impact of ENSO on tropical, lower-stratospheric ozone is robust and stable over time. Near 10 hPa the relationship between ozone and N3.4 is positive and approximately constant at ∼0.8%, regardless of the period examined. At these altitudes, the response of ozone arises from negative temperature anomalies associated with ENSO; lower temperatures reduce the loss ozone by the reaction O_3 + O, which is strongly temperature dependent. The negative temperature anomalies, in turn, are due to enhanced tropical upwelling during warm ENSO events.

[11] It should be noted that including N3.4 in the multiple regression for lower stratospheric ozone over 1979–2003 reduces the magnitude of the f10.7 coefficient by over one third (from 2.8% to 1.8% per 100 units), but the lower value is still more than twice as large as that of the regression

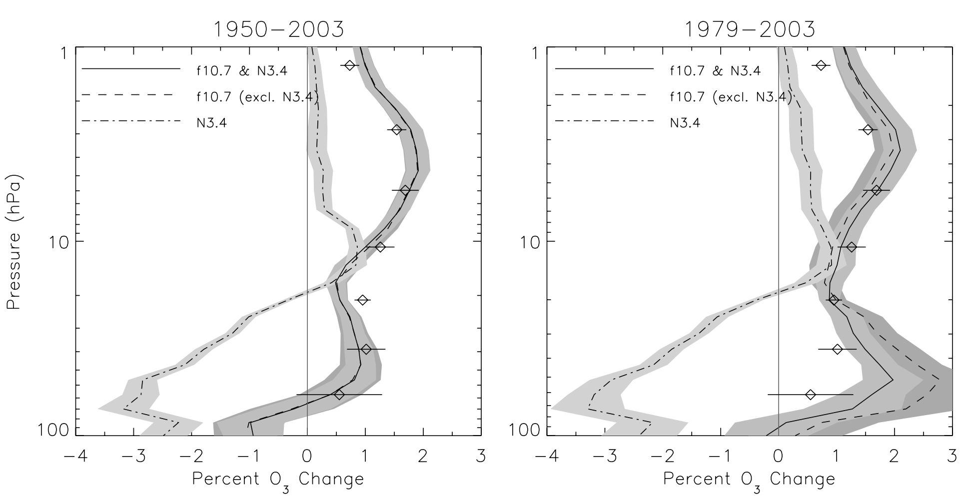

Figure 4. Vertical profiles of regression coefficients for tropical ozone (averaged over ±24°) for 1950–2003 and 1979–2003 determined from different multiple linear regressions. Profiles are shown for the coefficients of f10.7 (solid line) and N3.4 (dash-dotted line) when ozone is regressed on time, f10.7 and N3.4. Dashed line is the coefficient of f10.7 when the regression excludes N3.4. The coefficients are given as percent change per 100 units of f10.7 and per unit of N3.4; shaded areas indicate the 2σ uncertainty. For comparison, results from the fixed solar maximum vs. solar minimum calculations of *Marsh et al.* [2007] are shown as individual data points, with 2σ error bars, at selected levels.

coefficient obtained for the longer period 1950–2003 (0.74–0.78% per 100 units). Since, as shown above, the projection onto N3.4 does not change much regardless of the period analyzed, this implies either that inclusion of N3.4 cannot fully separate solar and ENSO induced variability over the shorter period, or that the response to the solar cycle is changing with time.

[12] For comparison we also show in Figure 4 the solar cycle response from calculations using fixed solar maximum vs. minimum forcing from the study of *Marsh et al.* [2007] at a few pressure levels. The response is very similar to the transient response for the period 1950 to 2003. This suggests that it may be necessary to have satellite data over one or two additional solar cycles in order to obtain stable observational estimates of the response of lower-stratospheric ozone to solar variability. Note also that the WACCM3 simulations discussed here do not include a QBO. In a model with a QBO, or indeed in the real atmosphere, the period necessary to obtain unambiguous estimates may be even longer.

[13] Finally, since much of the total ozone column, even in the Tropics, is contained within the lower stratosphere, ENSO variability should also be considered when investigating the solar signal in column ozone. Regression analysis on the ozone column was performed for the same periods and fit parameters as shown in Figure 4. A solar signal of 3.56 ± 0.16 Dobson Units (DU) per 100 units f10.7 was calculated if only the period 1979–2003 is considered and the N3.4 proxy is not included in the linear regression; the response decreases to 3.23 ± 0.13 DU if lagged N3.4 is included in the regression, and is reduced further to 2.25 ± 0.10 DU if the period 1950–2003 is analyzed. Only the first of these numbers (3.56 DU) is comparable to observational determinations from satellite data for the period 1979–2005 discussed by *Randel and Wu* [2007].

[14] It is also worth noting that, when ozone is regressed only on time and f10.7 for 1979–2003, the column change over the solar cycle in the Tropics is dominated by the lower stratosphere (below 20 hPa), which contributes 63% of the total; this is consistent with, albeit smaller than, the estimate of *Hood* [1997], who determined from satellite data that the lower stratosphere contributed 85% of the total column change over the solar cycle. However, when N3.4 is included as a predictor in our regression for 1979–2003, the contribution of the lower stratosphere to the column change over the solar cycle drops to 50%; and when the extended period 1950–2003 is analyzed, the lower stratosphere contributes only 39% of the total column change.

4. Discussion

[15] The analysis presented here suggests that a significant fraction of the decadal variability in satellite observations of ozone in the tropical lower stratosphere, which has been attributed to the effect of the solar irradiance cycle, may in fact be related to changes in tropical SST. Our model results show that variability in lower-stratospheric ozone is strongly related to changes in tropical upwelling associated with ENSO, and that inclusion of the N3.4 ENSO proxy, with a suitable time lag, in multiple linear regressions reduces considerably the apparent solar signal in ozone below about 20 hPa. We find that a lag of approximately 5 months produces the strongest projection onto N3.4 when multiple regression is carried out.

[16] *Randel and Wu* [2007] found large differences between ground-based and satellite estimates of the solar response of total ozone. The ground-based response in the tropics was approximately 2 DU less per 100 units f10.7 than the response determined from SBUV satellite data. Considering that the ground-based observations extend over the period from 1964 to the present, while SBUV observations began in 1979, this discrepancy appears to be consistent with the model results presented above. That is, it seems likely that part of the SBUV total tropical column changes are related to ENSO and not the solar cycle.

[17] We note in closing that, while other modeling studies have addressed the impact of SST on variability in the middle atmosphere, they have emphasized changes in the extratropical circulation and polar ozone [e.g., *Braesicke and Pyle*, 2004]. However, a recent simulation of the response of the atmosphere to doubled CO_2 by *Fomichev et al.* [2007] shows that upwelling in the tropical lower stratosphere strengthens (and ozone mixing ratios decrease) as SST increase under $2 \times CO_2$ conditions. This is consistent with our results for SST increases associated with warm ENSO events, and suggests that the same mechanism (the details of which are beyond the scope of the present paper) operates in both cases.

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R. R. Garcia and D. R. Marsh, National Center for Atmospheric Research, P.O. Box 3000, Boulder, CO 80307-3000, USA. (rgarcia@ucar.edu; marsh@ucar.edu)