

Increases in global tropospheric ozone following an El Niño event: examining stratospheric ozone variability as a potential driver

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Abstract

The stratosphere can strongly influence the interannual variability of tropospheric ozone. It has been discussed previously that tropospheric ozone can increase following an El Niño event, due to enhanced stratosphere/troposphere exchange (STE) of ozone. Here, we run a chemical-transport model for 5 years, covering a period including a strong El Niño event (1997–1998), and find that variability of ozone in the stratosphere is an almost negligible driver of modelled post-El Niño increases of ozone STE and tropospheric ozone abundances. Changes in the dynamics, affecting the cross-tropopause air-mass flux, may be far more important in driving these anomalies. Copyright © 2011 Royal Meteorological Society

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1. Introduction

Tropospheric ozone is a major pollutant affecting human health and influencing global and regional climate. Its concentration is controlled by a complex range of chemical reactions that influence production and destruction. Ozone is also deposited on the Earth's surface, while the troposphere is a net importer of ozone from the stratosphere, a region where ozone abundances are large. Stratosphere/troposphere exchange (STE) of ozone (hereafter, STEO3) occurs regularly, with a maximum during spring, and its contribution to tropospheric ozone concentrations has been discussed in various observational studies, often by correlating lower stratospheric with tropospheric ozone values (e.g. Ordóñez *et al.*, 2007; Terao *et al.*, 2008). Overall, STEO3 is still a major uncertainty in current modelling of the global tropospheric ozone budget (Wild, 2007; Wu *et al.*, 2007). The fact that the majority of models (e.g. Collins *et al.*, 2003; Sudo *et al.*, 2003; Hegglin and Shepherd, 2009) project a higher contribution of STEO3 to future tropospheric ozone, due to an enhancement of the Brewer–Dobson circulation and due to stratospheric ozone recovery (Zeng *et al.*, 2010), makes understanding of the processes that control STEO3 important.

STEO3 variability can be affected both by dynamical processes (residual mean circulation, isentropic mixing associated with tropopause folds) and by variations in stratospheric ozone abundances (Stohl *et al.*, 2003; Salby and Callaghan, 2006). Here, we focus on interannual variability, as influenced by an El Niño event. El Niño is known to be a major driver of anomalies in tropical tropospheric chemistry (e.g.

Chandra *et al.*, 1998; Doherty *et al.*, 2006). Its effects on extratropical tropospheric chemistry have been investigated only during the past few years. Zeng and Pyle (2005) examined the influence of El Niño on modelled STEO3 and global tropospheric ozone. They found that El Niño can be followed by positive global tropospheric ozone anomalies, due to increases in STEO3, which could have been caused by modulations of the subtropical jet (Shapiro *et al.*, 2001). The result was consistent with the previous observational finding of Langford *et al.* (1998). Koumoutsaris *et al.* (2008) thoroughly examined the impact of El Niño on northern extratropical tropospheric ozone using a different model and reached to a similar conclusion about the effect on STEO3. This interaction between climate and composition could be of more importance in the future, if El Niño events become more frequent/intense. However, the impact of climate change on El Niño remains fairly uncertain (Collins *et al.*, 2010).

These previous studies did not examine whether the STEO3 anomalies during El Niño are associated with stratospheric ozone changes. It is important to understand the role of these interactions in the past and, hence, the possible implications for a future atmosphere. Here, we use a chemical-transport model (CTM) to investigate whether anomalies in stratospheric ozone are an important driver of the increases in STEO3 and tropospheric ozone following El Niño. Our study does not aim to analyse in detail what drives the dynamical or the stratospheric ozone anomalies, but focuses on the possible impacts of such anomalies on tropospheric ozone.

2. Model description and experimental set-up

We use the Cambridge *p*-TOMCAT tropospheric CTM [a different model from what was used in Zeng and Pyle (2005)] as described and validated in Voulgarakis *et al.* (2009a). The CTM accounts for 79 gaseous chemical species and uses offline 6-hourly varying meteorology from the European Center for Medium range Weather Forecasting (ECMWF) operational analyses to drive the chemistry. Here, for the top model levels (above 100 hPa), ozone was relaxed to 5-day varying values for 1996–2000, obtained from the SLIMCAT CTM (for SLIMCAT validation see Chipperfield, 1999; for the experimental set-up of the runs which produced the data used here see Hadjinicolaou *et al.*, 2005). Since SLIMCAT is a stratospheric model, this relaxation will provide a more realistic representation of spatial and temporal variabilities of stratospheric ozone in *p*-TOMCAT. In Hadjinicolaou *et al.* (2005), it was demonstrated that SLIMCAT has a good skill in simulating stratospheric ozone variability at middle latitudes. Also, Voulgarakis *et al.* (2010) showed that *p*-TOMCAT captures the year-to-year changes of tropospheric ozone in 1996–2000, as observed at various sites around the globe.

Two experimental runs were performed: (1) BASE, in which no perturbation was applied, and (2) StratO3Fix, in which ozone above 100 hPa was kept fixed at 1996 values (still varying every 5 days). Specifying stratospheric ozone may affect its tropospheric abundances through both STEO3 changes and the modulation of modelled radiation, which impacts photolysis. Note that StratO3Fix does not, of course, represent a realistic state of the atmosphere, since many of the processes in the system are coupled. Stratospheric ozone interannual variability is linked to the interannual variability of stratospheric dynamics and the strength of the meridional Brewer–Dobson circulation, which, in turn, is the driver of the strength of STE (Holton *et al.*, 1995; Appenzeller *et al.*, 1996; Stohl *et al.*, 2003; Salby and Callaghan, 2006). However, our sensitivity approach can provide valuable information on the factors controlling tropospheric ozone variability, following a similar approach to Voulgarakis *et al.* (2010). The latter examined the drivers of interannual variability of tropospheric tracers, through the analysis of sensitivity runs in which each factor affecting variability (emissions, meteorology, clouds) was ‘fixed’ to not vary interannually.

The model was run from June to December 1995 for spin-up and the years analysed were 1996–2000. This period includes an exceptionally strong El Niño event, which commenced around May 1997, peaked in December 1997 and ended around June 1998 (McPhaden, 1999). The horizontal resolution is $2.8^\circ \times 2.8^\circ$, with 31 vertical layers extending from the surface to 10 hPa. Note that when referring to STE or STEO3 in this article, we refer to the net downward

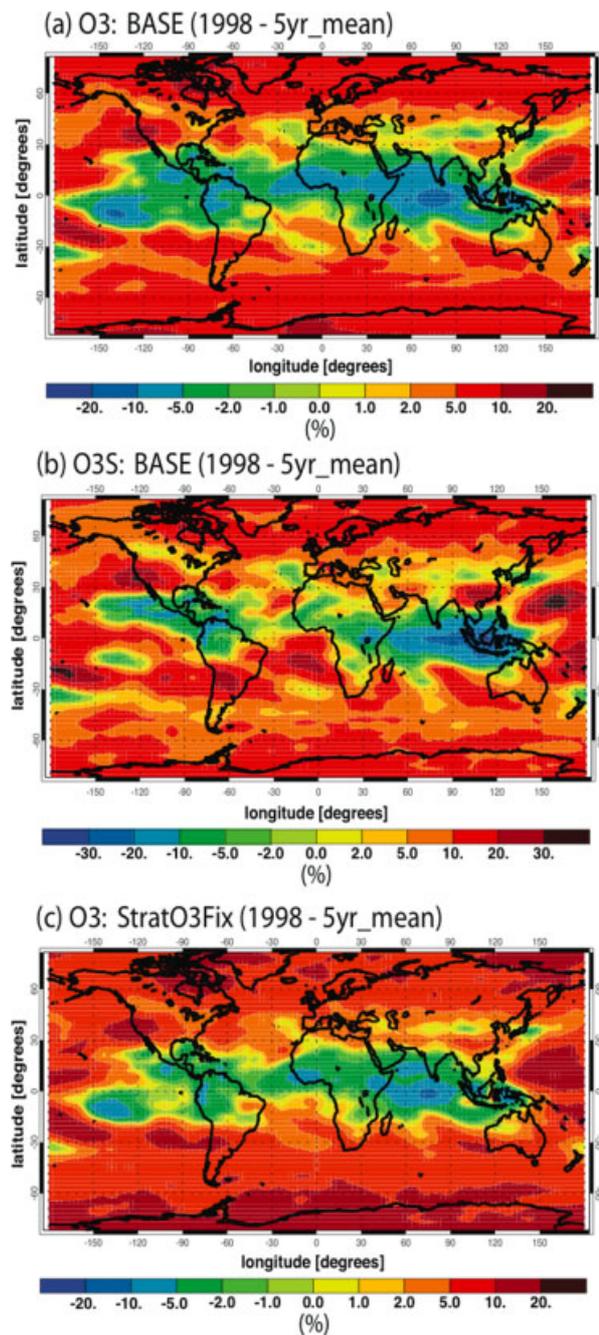


Figure 1. Percentage (%) differences from the 5-year mean of the 1998 annual mean a) tropospheric O_3 column in BASE, b) tropospheric O_3S column in BASE, and c) tropospheric O_3 column in StratO3Fix.

cross-tropopause flux (predominantly positive when summed over the extratropics or over the whole globe).

3. Results and discussion

Figure 1(a) shows the difference from the 5-year mean of the annual mean tropospheric ozone columns in 1998, the year following the peak of the strong El Niño event. It is clear that during this year tropospheric ozone increased in a very large part of the globe, especially in the extratropics. This is consistent with the findings of Zeng and Pyle (2005), Koumoutsaris *et al.*

(2008) and Voulgarakis *et al.* (2010), which used the UM/CHEM, GEOS-Chem and *p*-TOMCAT models, respectively.

Figure 1(b) shows the same as Figure 1(a) but for O₃S. O₃S is a ‘tagged’ stratospheric ozone tracer, which originates in the stratosphere and is affected by transport, deposition and chemical loss in the troposphere. In Figure 1(b), the geographical pattern of the O₃S differences for 1998 is very similar to the pattern of tropospheric ozone differences (Figure 1(a)). This implies that the extratropical increases of tropospheric ozone were influenced by the increased presence of stratospheric ozone-rich air in the troposphere. Another possibility could have been that decreased ozone chemical loss in the troposphere drives the ozone/O₃S correlation patterns, since loss processes are identical for tropospheric ozone and O₃S. However, the chemical loss of extratropical tropospheric ozone (even when normalized by the ozone burden, which partly influences the loss) is higher than average in 1998 (not shown), and thus cannot be the cause of the tropospheric ozone and O₃S increases. Therefore, in agreement with the previous studies, we suggest that the ubiquitous increases of extratropical tropospheric ozone in 1998 are mainly related to the input of ozone from the stratosphere.

In Figure 2, we show the results for the global ozone budget. Focusing on BASE run results (dark grey), it is shown that global tropospheric ozone is maximum (12 Tg higher than average) in 1997/1998. The high global values for 1998 are consistent with the extratropical features of Figure 1(a). Also, it is clear that the 1998 tropospheric ozone peak is not associated with increased global tropospheric net chemical tendency of ozone (NetChemO₃) during this year (Figure 2(c)). On the other hand, anomalously high global STEO₃ occurs in 1998 (Figure 2(b)), with around 100 Tg more ozone than average entering the troposphere from the stratosphere during this year.

In Figure 3, we show global STEO₃, net ozone chemical tendency (NetChemO₃) and the ozone burden, but as 12-month running monthly means and together with the Oceanic Niño Index (ONI). It can be seen that there is a roughly 6-month lag between the peak of El Niño and the maximum/minimum of STEO₃/net chemistry. This is in agreement with Zeng and Pyle (2005) and Koumoutsaris *et al.* (2008) (for STEO₃). It is also clear that STEO₃ and NetChemO₃ are very well anti-correlated, which indicates that what drives the NetChemO₃ decrease in 1997–1998 is the increase of ozone chemical loss, as a consequence of increases in STEO₃. Note that the ozone production (not shown) does not follow a similar evolution in 1996–2000, but just has a profound maximum in 1997, when high amounts of biomass burning pollution were present in the tropical troposphere. This maximum does not affect the evolution of NetChemO₃ drastically. It is also interesting to note that 6 months after the ONI becomes negative, STEO₃ starts to decrease. La Niña conditions (which progressively started in

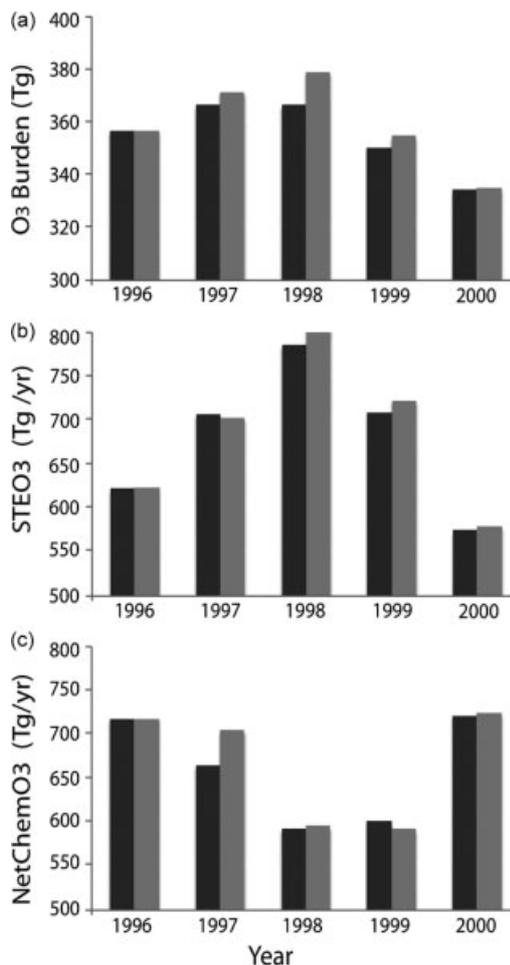


Figure 2. Year-to-year variations of annual mean a) tropospheric ozone burden, b) STEO₃, and c) tropospheric ozone net chemical tendency (NetChemO₃), in the BASE (dark grey) and the StratO₃Fix (light grey) runs.

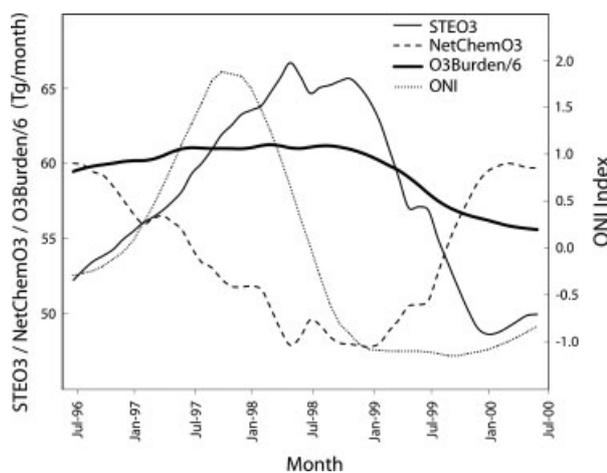


Figure 3. 12-month running monthly means of STEO₃ (solid line), ozone net chemical tendency (NetChemO₃; dashed line) and ozone burden (thick solid line). The latter is divided by 6 to be on the same scale as the other quantities. The ONI (Oceanic Niño Index: 3-month running mean of ERSST.v3b SST anomalies in the Niño 3.4 region (5N–5S, 120–170W)) is plotted as well (dotted line; right axis), in order to interpret the ozone results.

summer 1998) have the opposite effects on global tropospheric ozone to those of El Niño.

What is particularly interesting in Figure 2 is that the terms of the annual tropospheric ozone budget seem not to be strongly affected when fixing stratospheric ozone to 1996 values (comparison of light and dark grey bars). In particular, the STEO3 variability is hardly different when we exclude year-to-year changes in stratospheric ozone abundances. Thus, it is clear that the anomalously high STEO3 in 1998 (as well as the low STEO3 values of 2000) in the model are not driven by changes in stratospheric ozone, and they can only be linked to variability in the cross-tropopause air-mass flux. We reach similar conclusions by comparing Figure 1(c) (StratO3Fix) with Figure 1(a) (BASE): in most parts of the globe, tropospheric ozone is not affected by the interannual variability of stratospheric ozone.

There exists a possibility that an underestimation of the variability of stratospheric ozone during the period of study by the model could have led to false conclusions about the importance of this driver. However, when examining the standard deviation (indicating variability) of the total ozone column (dominated by the stratospheric component) in our model *versus* TOMS/SBUV observations (Table I), we find that for all but one of the seasons and regions examined, the variability in the model is either similar to the variability in the observations or larger than the latter. Similar or larger variability in the model would not imply an underestimation of the effect of stratospheric ozone changes on the annual mean tropospheric ozone budget.

Note that in 1998, both in the northern and in the southern extratropics (30°N–50°N; 30°S–50°S), total ozone columns in the model are higher than the 1996–2000 average (not shown), as they also are in the observations, but with anomalies being less than +2% in both hemispheres. Anomalies of lower stratospheric (between 100 hPa and the tropopause) model ozone in 1998 are around +3 to 7%, both for the globe and for the extratropics separately, in good agreement with Zeng and Pyle (2005). Note that these anomalies remain very similar in the StratO3Fix run (not shown), implying that in both studies this

Table I. Standard deviation (indicating variability) of total ozone column (in DU) in the TOMS/SBUV observations (OBS) and in the model (MOD).

	JFM (OBS/ MOD)	AMJ (OBS/ MOD)	JAS (OBS/ MOD)	OND (OBS/ MOD)
30N, 50N	20.6/24.2	12.5/19.0	8.0/18.0	9.4/25.6
30S, 30N	8.6/8.6	6.4/7.6	8.9/4.0	8.2/8.2
30S, 50S	4.6/9.2	10.4/10.2	21.1/23.3	6.3/11.9

The values shown are calculated using seasonal averages [January–February–March (JFM), April–May–June (AMJ), July–August–September (JAS), October–November–December (OND)] from 5 years (1996–2000) of data. Regions examined are the northern midlatitudes (30N–50N), the tropics (30S–30N) and the southern midlatitudes (30S–50N). Values are not shown for north of 50°N or for south of 50°S, as data coverage is not full at these latitudes. For details about the TOMS/SBUV observations used, see Frith *et al.* (2004)

lower stratospheric ozone anomaly is part of the broader feature of increased downward flux of ozone as a consequence of El Niño, and is not sensitive to stratospheric ozone variability higher up. On the other hand, global positive STEO3 anomalies in 1998 are as high as 16% (Figure 2(b)). The above enhance our confidence in the conclusion that stratospheric ozone anomalies are most likely not the main driver of the post-El Niño tropospheric ozone increases. Although here both observed and modelled total ozone columns are slightly higher than average following the 1998 El Niño, in qualitative agreement with Brönnimann *et al.* (2004) who studied the 1940–1942 El Niño, an earlier study (e.g. Zerefos *et al.*, 1992) concluded that it is more likely that El Niño has a slightly negative impact on extratropical total ozone. The possible links between El Niño and stratospheric ozone variability need to be further investigated.

Finally, note that the other potential contribution of stratospheric ozone changes to tropospheric ozone variability, photolysis modification, apparently is also small in the model. Had there been a major influence through this path, the net chemistry of tropospheric ozone in Figure 2(c) would have been different between BASE and StratO3Fix.

4. Conclusions and discussion

We have examined the effect that year-to-year changes in stratospheric ozone abundances have on tropospheric ozone concentrations, during a period (1996–2000) dominated by an anomalous El Niño event. Past studies demonstrated that El Niño can drive increases in the amounts of stratospheric ozone entering the troposphere through STE. These findings are confirmed here. We also find that the variability of the STEO3 remains almost unaffected when ignoring stratospheric ozone interannual variability in the model. This indicates that the large increases in modelled STEO3 in 1998 (following El Niño) are not associated with the interannual variability in stratospheric ozone abundances and are almost solely caused by anomalies in atmospheric transport processes driving the cross-tropopause air-mass flux. Our results also point out that caution is needed when attributing tropospheric ozone changes to STEO3 by correlating low stratospheric with tropospheric ozone concentrations. Investigation of the role of the dynamics is warranted.

The fact that during the period of study stratospheric ozone changes do not appear to be a driver of changes in tropospheric ozone does not imply that this driver cannot be important in other cases. There could be years when interannual variability in cross-tropopause transport processes is small, but when stratospheric ozone reaches highly anomalous concentrations. In this case, stratospheric ozone variability could be a non-negligible driver. Furthermore, variability of stratospheric ozone can impact tropospheric ozone through modifications of photolysis

(e.g. Zanis *et al.*, 2002; Isaksen *et al.*, 2005), especially on local/regional scales, which can also be important for other key tropospheric species (e.g. OH; see Voulgarakis *et al.*, 2009b), even in years when the influence of transport processes is overwhelming for global tropospheric ozone. The effect of stratospheric ozone changes on global and regional tropospheric oxidation via photolysis modifications should be the subject of a further study.

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