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To cite this version:
Daniel Cariolle, M.J. Evans, M.P. Chipperfield, N. Butkovskaya, Alexandre Kukui, et al.. Impact of the new HNO3-forming channel of the HO2+NO reaction on tropospheric HNO3, NOx, HOx and ozone. Atmospheric Chemistry and Physics, European Geosciences Union, 2008, 8 (14), pp.4061-4068. 10.5194/acp-8-4061-2008 . hal-00309118

HAL Id: hal-00309118
https://hal.archives-ouvertes.fr/hal-00309118
Submitted on 8 Jan 2016

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Impact of the new HNO$_3$-forming channel of the HO$_2$+NO reaction on tropospheric HNO$_3$, NO$_x$, HO$_x$ and ozone

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Received: 4 December 2007 – Published in Atmos. Chem. Phys. Discuss.: 11 February 2008
Revised: 24 June 2008 – Accepted: 2 July 2008 – Published: 25 July 2008

Abstract. We have studied the impact of the recently observed reaction NO+HO$_2$→HNO$_3$ on atmospheric chemistry. A pressure and temperature-dependent parameterisation of this minor channel of the NO+HO$_2$→NO$_2$+OH reaction has been included in both a 2-D stratosphere-troposphere model and a 3-D tropospheric chemical transport model (CTM).

Significant effects on the nitrogen species and hydroxyl radical concentrations are found throughout the troposphere, with the largest percentage changes occurring in the tropical upper troposphere (UT). Including the reaction leads to a reduction in NO$_x$ everywhere in the troposphere, with the largest decrease of 25% in the tropical and Southern Hemisphere UT. The tropical UT also has a corresponding large increase in HNO$_3$ of 25%. OH decreases throughout the troposphere with the largest reduction of over 20% in the tropical UT. The mean global decrease in OH is around 13%, which is very large compared to the impact that typical photochemical revisions have on this modelled quantity. This OH decrease leads to an increase in CH$_4$ lifetime of 5%. Due to the impact of decreased NO$_x$ on the OH:HO$_2$ partitioning, modelled HO$_2$ actually increases in the tropical UT on including the new reaction. The impact on tropospheric ozone is a decrease in the range 5 to 12%, with the largest impact in the tropics and Southern Hemisphere. Comparison with observations shows that in the region of largest changes, i.e. the tropical UT, the inclusion of the new reaction tends to degrade the model agreement. Elsewhere the model comparisons are not able to critically assess the impact of including this reaction. Only small changes are calculated in the minor species distributions in the stratosphere.

1 Introduction

It is well established that the reaction of HO$_2$ with NO plays a central role in atmospheric chemistry:

$$\text{NO} + \text{HO}_2 \rightarrow \text{OH} + \text{NO}_2 \quad (R1)$$

In the stratosphere, this reaction moderates the effectiveness of the cycle involving HO$_x$ (OH, HO$_2$) radicals that is an important removal mechanism of ozone (see e.g. Wayne, 2000). In the troposphere Reaction (R1) plays a key role in controlling the interconversion between HO$_2$ and OH radicals through cycles involving the reactions:

$$\text{OH} + \text{CO}(+\text{O}_2) \rightarrow \text{HO}_2 + \text{CO}_2 \quad (R2)$$

$$\text{OH} + \text{VOCs}(+\text{O}_2) \rightarrow \text{HO}_2 + \text{products} \quad (R3)$$

The VOCs include methane, non-methane hydrocarbons and other volatile carbon-containing species. Reaction (R1) is a secondary source of the OH radical (the major tropospheric oxidant), as well as the major source of tropospheric ozone through the conversion of NO to NO$_2$ followed by NO$_2$ photolysis. The OH and ozone production rates are limited mainly by the chain termination reaction:
Fig. 1. Pressure and temperature dependences of $\beta = k_{\text{R1b}}/k_{\text{R1}}$ in percent. Upper and lower dotted lines represent extrapolation to $P=760$ and $P=50$ Torr, respectively. The arrow corresponds to typical $\beta$ values as a function of altitude from the Earth’s surface to the tropopause region.

\[ \text{OH + NO}_2 + \text{M} \rightarrow \text{HNO}_3 + \text{M} \quad \text{(R4)} \]

Another potential significant chain termination reaction is the minor HNO$_3$-forming channel of the reaction of HO$_2$ with NO that has been observed in laboratory experiments by Butkovskaya et al. (2005):

\[ \text{HO}_2 + \text{NO} \rightarrow \text{HNO}_3 \quad \text{(R1b)} \]

The branching ratio, or rate constant ratio, $\beta = k_{\text{R1b}}/k_{\text{R1}}$, for the new Reaction (R1b) was found to range from $\sim 0.2$ to $0.8\%$ from $300\,\text{K}$ to $200\,\text{K}$, at a pressure of $200\,\text{Torr}$. These first data led to the suggestion that Reaction (R1b) could be a significant loss process of HO$_3$ radicals in the upper troposphere. To assess the potential importance of Reaction (R1b) throughout the troposphere, it appeared necessary to determine the branching ratio $\beta$ over the full range of tropospheric pressures and temperatures in order to provide a parametrisation of $\beta$ for atmospheric modelling.

In this paper we first briefly report the experimental measurements of $\beta$ as a function of pressure and temperature in the ranges 70–600 Torr and 220–320 K. Then, we present 2-D and 3-D model calculations of the impact of Reaction (R1b) on the tropospheric abundances of HNO$_3$, HO$_3$, NO$_3$ and ozone by including the parametrisation equation of $\beta$ obtained from the laboratory experiments. Finally, the model results are compared with observations.

2 Laboratory experiments

The HNO$_3$-forming channel of the HO$_2$+NO reaction has been investigated using a turbulent flow reactor (TFR) coupled to a chemical ionisation mass spectrometer (CIMS). The experimental set up has previously been described in detail (Butkovskaya et al., 2005). Briefly, the required high flowrates (60 to 150 SLPM) are obtained by flowing N$_2$ carrier gas from a liquid nitrogen tank. The working pressure and temperature ranges in the TFR are 70–600 Torr and 220–320 K, respectively. The HO$_2$ radicals were produced by reaction of H atoms with O$_2$. The species of interest in the reactive system, HO$_2$, OH, NO$_2$ and HNO$_3$, were detected by CIMS from their reaction with the SF$_6^+$ ion. The branching ratio of Reaction (R1), $\beta = k_{\text{R1b}}/k_{\text{R1}}$, was obtained by measuring the concentration ratio of the HNO$_3$ and NO$_2$ products from channels (R1b) and (R1), respectively.

The calibration procedure of the CIMS signals of HNO$_3$ and NO$_2$ is described in detail by Butkovskaya et al. (2007). In order to increase the signal intensity of produced HNO$_3$, which was a critical factor in reducing experimental uncertainties, the signal was chemically amplified by adding high concentrations of CO and O$_2$ into the TFR. The chain mechanism (Reactions R1 and R2) that occurred in the TFR is similar to that occurring in chemical amplifiers used to measure peroxy radicals in the atmosphere (e.g. Cantrell and Stedman, 1982). In addition, CO in Reaction (R2) acts as a scavenger of OH produced in Reaction (R1), preventing formation of HNO$_3$ in the secondary reaction of OH with NO$_2$ (Reaction R4). The HNO$_3$ formation in the OH + NO$_2$ reaction contributed up to 10% of the yield from HO$_2$ + NO reaction at typical concentrations of CO, O$_2$ and NO used in these experiments, as also calculated by numerical simulation of the kinetics in the reactor (see Butkovskaya et al., 2007 for details).

The branching ratio $\beta$ was measured at different pressures and temperatures in the ranges 70–600 Torr and 223–323 K. The pressure and temperature dependencies of $\beta$ are shown in Fig. 1. In the 298–223 K range the temperature dependences for different pressures present a set of nearly parallel straight lines. The linear fit fails at temperatures higher than 298 K, where curvatures were observed for $P=100$, 200 and 400 Torr. Below 298 K the whole set of data can be described by the simple three-parameter expression of the general form:

\[ \beta(P, T) = a/T + bP + c \quad (1) \]

Coefficient $a$ was found by averaging the slopes of the observed temperature dependencies. Coefficients $b$ and $c$ were determined by standard two-parameter least-square fit of the data. The numerical expression of $\beta$ in percent can be written as:

\[ \beta(P, T) = k_{\text{R1b}}/k_{\text{R1}} = (530 \pm 20)/T (\text{K}) + (6.4 \pm 1.3) \times 10^{-4} P(\text{Torr}) - (1.73 \pm 0.07) \quad (2) \]

(with 2σ uncertainties). The rate constant $k_{\text{R1b}}$ has been derived from this equation by considering the recommended value $k_{\text{R1}}=3.5 \times 10^{-12}\text{exp}(250/T)\text{cm}^3\text{molecule}^{-1}\text{s}^{-1}$ (Sander et al., 2006). This NASA-JPL evaluation panel
recommended value and the one recommended by the IUPAC panel (Atkinson et al., 2004, updated 2006) are very similar, the IUPAC one being higher by only 14% at 200 K and 10% at 300 K.

3 Model calculations

We have used two atmospheric chemical models to investigate the impact of Reaction (R1b) on atmospheric chemistry. Using the 2-D (latitude-height) stratosphere-troposphere MOBIDIC model (Cariolle and Brand, 1984) it is possible to evaluate the significance of Reaction (R1b) on the global chemical cycles. It is found that this reaction is important in the free troposphere to control the NO\textsubscript{3} and HO\textsubscript{x} radical contents and the ozone concentration at low latitudes.

We have also used the GEOS-CHEM 3-D tropospheric chemical transport model (e.g. Evans et al., 2005). This model has a more detailed treatment of tropospheric chemical processes and allows an assessment of the impact of Reaction (R1b) under a wider range of conditions and a more critical comparison against tropospheric observations.

3.1 Model simulations and impact on species distributions

The MOBIDIC model has been recently updated to include the Sander et al. (2006) compilation for the reaction rates of atmospheric interest. The dynamical forcing and the numerical scheme for the transport of minor trace species have also been updated (Cariolle and Teyssédré, 2007).

To investigate the impact of Reaction (R1b) two simulations have been performed. The control simulation (referred to as run CON2D) does not include Reaction (R1b) while simulation NEW2D includes Reaction (R1b) with \( k_{R1b} = \beta \times k_{R1} \) and \( \beta \) given by Eq. (2). Each simulation was integrated over 8 years. The model outputs are compared for the last simulated year when steady-state is reached.

The GEOS-CHEM 3-D tropospheric CTM has been used for a number of studies focused on tropospheric O\textsubscript{3}-OH-NO\textsubscript{x} budgets (e.g. Bey et al., 2001; Evans and Jacob, 2005 and references therein). The GEOS-CHEM model is driven by assimilated winds calculated by the Goddard Earth Observing System at 4\(^\circ\) x 5\(^\circ\) horizontal resolution and 30 levels in the vertical. Surface sources of NO\textsubscript{x} are 42 Tg N yr\(^{-1}\) (anthropogenic, biomass burning and soil) while upper troposphere sources amount to 4 Tg N yr\(^{-1}\) (lightning and aircraft). The model description is given by Bey et al. (2001).

In this study we have used the same model version as Evans and Jacob (2005) who investigated the impact of new laboratory studies on the aerosol hydrolysis of N\textsubscript{2}O\textsubscript{5}. In that study Evans and Jacob found that the newer slower rate of conversion of N\textsubscript{2}O\textsubscript{5} to HNO\textsubscript{3} on aerosols increased the average tropospheric budgets of NO\textsubscript{x} (by 7%), O\textsubscript{3} (4%), and OH (8%), bringing the model into better agreement with observations. Here we repeat the approach of Evans and Jacob (2005) and include Reaction (R1b).

The 3-D model was run for 2 simulations of 2 years each. The control run (CON3D) was similar to a run from Evans and Jacob (2005) using the aerosol parameters from their Table 1. Run NEW3D was the same as run CON3D but included Reaction (R1b) in the same way as for the MOBIDIC model. The first year of the model run was treated as a spin up and results from the second year were used for analysis.

3.1.1 Impact on nitrogen species

Figure 2a shows the rate of HNO\textsubscript{3} formation by Reaction (R1b) and by Reaction (R4) between NO\textsubscript{2} and OH at the equator in March using output from run NEW2D. It appears that Reaction (R1b) becomes a significant channel for HNO\textsubscript{3} formation between 7 to 16 km in the cold upper troposphere region. The ratio \( Ra \) between the rates of HNO\textsubscript{3} formation by these 2 reactions is equal to: \( Ra = ([NO_3],[HO_2]/[NO_2],[OH]).(k_{R1b}/k_{R4}). \) The first term of the product corresponds to an “atmospheric” component that depends on the NO\textsubscript{x} and HO\textsubscript{x} profile concentrations. This term increases from a value of 40 at ground level to a peak value of about 220 at 10 km. The second term \( k_{R1b}/k_{R4} \) shows an increase from 0.004 at ground level to 0.01 at 15 km. The differences in the altitude profiles of NO\textsubscript{x} and HO\textsubscript{x} appears as important as the temperature dependency of the Reaction (R1b) to determine its effectiveness. Both terms act in the same direction, so the HNO\textsubscript{3} concentration increases from ground to the UT when Reaction (R1b) is introduced in the 2-D model.

Figure 3 shows the variations in the HNO\textsubscript{3} concentration between 2-D model runs NEW2D and CON2D for the month of March. The relative effect is largest at the cold equatorial tropopause with an increase over 30%. HNO\textsubscript{3} increases by about 10% in the lower stratosphere and by up to more than 100% in the upper stratosphere. However, the HNO\textsubscript{3} concentrations are quite small compared to those of the other NO\textsubscript{x} species, so we expect little impact of Reaction (R1b) at those altitudes.

The increase in the HNO\textsubscript{3} concentration by Reaction (R1b) changes the ratio between NO\textsubscript{x} and HNO\textsubscript{3}. Model results show that the concentrations of NO\textsubscript{x} decreases by 20 to 30% in the troposphere. In particular Fig. 3 shows that the NO concentration decreases by more than 30% at low latitudes in the middle-upper troposphere.

Equally, Fig. 4 shows the zonal mean differences between runs CON3D and NEW3D for selected species in March. In run NEW3D NO\textsubscript{x} decreases everywhere ranging from –5% in the Northern Hemisphere (NH) mid-latitudes to –35% in the tropical and Southern Hemisphere (SH) upper troposphere (UT). The 3-D model shows more latitudinal variation in the modelled NO\textsubscript{x} response than the 2-D model. In the 3-D model, near the surface, the smallest relative NO\textsubscript{x} decrease occurs in the northern mid-latitudes in regions of high NO\textsubscript{x}.
close to emissions sources. In the Southern Hemisphere, away from fresh pollution, the long transport timescales for NO\textsubscript{x} allow a larger impact of Reaction (R1b). The 2-D model, with its simpler treatment of the troposphere, misses this effect partly because of specification of fixed surface mixing ratios in the 2 runs. Except very near the surface, HNO\textsubscript{3} increases everywhere in the 3-D model and by up to 35\% in the tropical UT. Overall, the MOBIDIC and GEOSCHEM models exhibit similar responses to the introduction of reaction (R1b).

3.1.2 Impact on HO\textsubscript{x} species

These perturbations in the nitrogen species budget impact the distribution of HO\textsubscript{x}. Figure 2b shows the rates of the main reactions that control the OH:HO\textsubscript{2} ratio. As can be seen, Reaction (R1) is a major player controlling this ratio in the middle troposphere – lower stratosphere at the equator (5–20 km). Consequently, the tropospheric NO decrease due to Reaction (R1b) tends to slow down the conversion of HO\textsubscript{2} to OH by Reaction (R1).

The global tropospheric OH budget is difficult to establish. It results from a delicate balance between production processes dominated by water vapour, methane and acetone (e.g. Arnold et al., 2004) decompositions and destruction by reactions involving HO\textsubscript{x} radicals that form back water vapour. As discussed by Butkovskaya et al. (2005) an additional effect of Reaction (R1b) is to decrease the HO\textsubscript{x} concentrations. Larger HNO\textsubscript{3} concentrations are formed that are subsequently washed out, thus removing a larger part of the HO\textsubscript{x} radicals. In addition, the HO\textsubscript{x} removal by the reaction between OH and HNO\textsubscript{3} can also be enhanced.

The net effect in the MOBIDIC model simulations is a decrease of the OH concentration by about 10 to 20\% in the middle and upper troposphere for latitudes lower than 40\° (Fig. 3), and a smaller increase of the HO\textsubscript{2} concentration by 10\% in the upper troposphere at the same latitudes between 10 to 15 km.

The GEOS-CHEM model shows a response consistent with the MOBIDIC model with however a larger contrast between the two hemispheres. The OH concentration decreases by −6\% at the surface at NH mid-latitudes, by −15\% at SH mid-latitudes, but as much as −20\% in the tropical UT. However, the decrease in NO\textsubscript{x} modifies the OH:HO\textsubscript{2} ratio and tends to increase the HO\textsubscript{2} concentration in some locations, e.g. the tropical UT, even though the overall net HO\textsubscript{x} production is reduced.

In run CON3D the mass-weighted global annual mean tropospheric OH is 1.12×10\textsuperscript{6} cm\textsuperscript{-3} and becomes 0.97×10\textsuperscript{6} cm\textsuperscript{-3} in run NEW3D, a decrease of 13\%. This is a very large relative change as modelled mean OH is a quantity that is typically well buffered against changes as revisions to model HO\textsubscript{x} photochemistry are implemented. However, this large relative change still leaves the model...
within the range of mean OH values derived from methyl-chloroform observations: 1.07±0.09/−0.17×10⁶ (Krol et al., 1998), 1.16±0.17×10⁶ (Srivastava et al., 2000) and 0.94±0.13×10⁶ cm⁻³ (Prinn et al., 2001). Those figures are consistent with recent evaluations by Bloss et al. (2005) that give a mean northern and southern hemispheric OH concentrations of 0.91 ×10⁶ cm⁻³ and 1.03 ×10⁶ cm⁻³ respectively, evaluations in very good agreement with the mean value in NEW3D.

The OH decrease will slow the destruction rates of important long-lived species which are sources of radicals or have large global warming potentials. For instance, CH₄ contributes significantly to the atmospheric greenhouse forcing. The major sink of CH₄ is its reaction with OH:

\[
\text{CH}_4 + \text{OH} \rightarrow \text{CH}_3 + \text{H}_2\text{O} \quad \text{(R5)}
\]

The CH₄ chemical lifetime, \( \tau_{\text{CH}_4} \), can therefore be evaluated by dividing its total atmospheric content by the integration over the whole domain of its destruction rate:

\[
\tau_{\text{CH}_4} = \frac{\int [\text{CH}_4] \, dv}{\int k_{\text{OH} + \text{CH}_4}[\text{OH}][\text{CH}_4] \, dv}
\]

Evaluations of \( \tau_{\text{CH}_4} \) from the MOBIDIC outputs give a lifetime of 9.6 years for run CON2D and 10.1 years for run NEW2D. Thus, the CH₄ lifetime is increased by about 5% when Reaction (R1b) is introduced in the model.

3.1.3 Impact on tropospheric ozone

These perturbations in the HOₓ and NOₓ distributions have an impact on the tropospheric O₃ budget. Ozone production via the cycles involving Reactions (R2) and (R3) is reduced due to the NO and OH decreases. The decrease calculated by MOBIDIC is largest at low latitudes between 5 and 10 km (Fig. 3), with a maximum of about \(-10\%\). This response is
compares annual mean modelled ozone profiles.

Figure 4. Mean differences (%) for the month of March between 3-D model runs NEW3D and CON3D for (a) NOx, (b) HO2, (c) HNO3, (d) CO, (e) O3 and (f) OH. Negative contours indicate smaller values in run NEW3D.

not very sensitive to the season. The GEOS-CHEM model gives a similar response in the UT, but produces a larger O3 decrease in the SH lower troposphere. This is due to the larger NOx decrease that is found in this hemisphere away from major surface sources. The decrease in NOx has the impact of decreasing O3 throughout the troposphere ranging from −6% to −16%.

3.2 Comparison with observations

Figure 5 compares modelled profiles of NOx and HNO3 from runs CON3D and NEW3D with a compilation of aircraft observations mapped onto a monthly 4° × 5° grid (Emmons et al., 2000). For this comparison the appropriate grid box of the model was sampled and then the results averaged over all the observation points. As discussed above, the NOx concentrations in run NEW3D are lower than run CON3D, while HNO3 is generally larger. The region where the modelled change is most significant is in the tropical UT where NEW3D agrees less well with the observations, especially for NOx. Elsewhere the modelled changes are not large compared with the model-observation differences.

Figure 6 compares annual mean modelled ozone profiles with the climatology of Logan (1999). In the middle troposphere both model runs agree fairly well with the observations, although run CON3D is, on average, closer. The model underestimate of observations of O3 in the 7–10 km range reflects issues with an overestimate in the GEOS tropopause mass flux which is in excess. To maintain a 500 Tg yr−1 cross tropopause O3 flux, lower than observed O3 concentrations are necessary within the model stratosphere. The mean ratio between model and observed in run CON3D is 1.0078 while in run NEW3D it is 0.9148. The total tropospheric O3 in run CON3D is 284 Tg while in run NEW3D it is 259 Tg. There are no direct observations of this total but these values are at the lower range of modelled values (see Stevenson et al., 2006).

4 Conclusions

The minor HNO3-forming channel of the reaction between NO and HO2 has a significant impact on tropospheric chemistry. It leads to a reduction of NOx and OH radical
This work was supported by the European Union FP6 Integrated Projects SCOUT-O3 (http://www.ozone-sec.ch.cam.ac.uk) and QUANTIFY (http://www.pa.op.dlr.de/quantify/), and the french LEFE/CHAT INSU programme (http://www.insu.cnrs.fr).

Acknowledgements. This work was supported by the

Fig. 5. Comparison of profiles (pptv) of NOx (top) and HNO3 (bottom) averaged over 3 latitude bands (left 90° S–30° S, centre 30° S–30° N, right 30° N–90° N) from 3-D model run CON3D (red line), run NEW3D (blue line) and observations (Emmons et al. (2000), black line).

concentrations that produces an ozone increase in the troposphere. Due to the temperature dependence of the reaction, the effects on the species distributions are generally largest at the equatorial mid and upper troposphere where the temperatures are low, and at high latitudes in the SH. In those regions the O3 concentration decreases by about 10% when Reaction (R1b) is taken into account.

The OH reduction induced by Reaction (R1b) can also increase the lifetime of the long-lived species whose destruction proceeds mainly by reaction with this radical. This is the case for CH4 whose lifetime reaches 10.1 years against 9.6 years when Reaction (R1b) is neglected.

Reaction (R1b) has been established by laboratory measurements and should therefore be included in photochemical models. Comparison of our 3-D model with observations of NOx, HNO3 and O3 suggests that the overall agreement tends to worsen when the reaction is included. This, therefore, points to other outstanding uncertainties in tropospheric NOx:HNO3 partitioning. Possible causes of these differences need further investigation in the laboratory and with models.

The present evaluation has been performed at global scale with a representation of the atmospheric chemistry adequate mostly for the free troposphere. It is expected that the impact of Reaction (R1b) could be different for specific regions where background species concentrations can vary from the mean values produced by the large-scale models. This, for instance, could be the case in boundary layers over continents or polluted areas. Thus, the present study should be complemented by air quality model simulations to obtain a more comprehensive picture of the possible regional and local influences of Reaction (R1b) on the atmospheric chemistry.

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