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Key Points:
- Large model overestimates of NO/NO_2 concentration ratios in the upper troposphere imply errors in $NO\text{-}NO_2\text{-}O_3$ cycling kinetics or the presence of an unaccounted labile NO_x reservoir
- The presence of an unaccounted labile NO_x reservoir would affect the NO_x lifetime in the upper troposphere and would suggest unrecognized, likely organic, chemistry
- Possible error in $NO\text{-}NO_2\text{-}O_3$ cycling kinetics would have large implications for global simulations of tropospheric ozone and for satellite retrievals of tropospheric NO_2

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Observed NO/NO_2 Ratios in the Upper Troposphere Imply Errors in $NO\text{-}NO_2\text{-}O_3$ Cycling Kinetics or an Unaccounted NO_x Reservoir

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Abstract Observations from the SEAC^{4}RS aircraft campaign over the southeast United States in August–September 2013 show NO/NO_2 concentration ratios in the upper troposphere that are approximately half of photochemical equilibrium values computed from Jet Propulsion Laboratory (JPL) kinetic data. One possible explanation is the presence of labile NO_x reservoir species, presumably organic, decomposing thermally to NO_2 in the instrument. The NO_2 instrument corrects for this artifact from known labile HNO_4 and $CH_3O_2NO_2$ NO_x reservoirs. To bridge the gap between measured and simulated NO_2, additional unaccounted labile NO_x reservoir species would have to be present at a mean concentration of ~40 ppt for the SEAC^{4}RS conditions (compared with 197 ppt for NO_x). An alternative explanation is error in the low-temperature rate constant for the $NO + O_3$ reaction (30% 1-σ uncertainty in JPL at 240 K) and/or in the spectroscopic data for NO_2 photolysis (20% 1-σ uncertainty). Resolving this discrepancy is important for understanding global budgets of tropospheric oxidants and for interpreting satellite observations of tropospheric NO_2 columns.

Plain Language Summary We identify large discrepancies between observed NO/NO_2 ratios and models representing our best understanding of the chemistry controlling NO and NO_2 in the upper troposphere over the southeast United States during August–September 2013. We suggest that either unrecognized chemistry or errors in modeled cycling between NO, NO_2, and O_3 could explain this discrepancy. Either explanation will have important implications for global tropospheric chemistry and for the interpretation of satellite observations of NO_2.

1. Introduction

Nitrogen oxide radicals ($NO_x \equiv NO + NO_2$) are emitted by anthropogenic sources (fuel combustion) and natural sources (lightning, soils, and fires). Anthropogenic emissions degrade surface air quality by producing ozone and nitrate particulate matter and also affect ecosystems through nitrogen deposition. On a global scale, NO_x increases the concentration of tropospheric oxidants (ozone and OH) with complicated implications for climate forcing (Wild et al., 2001). NO_x in the upper troposphere and the associated cycling between NO and NO_2 are of particular importance for production of tropospheric ozone and OH (Murray et al., 2013; Newsome & Evans, 2017; Ridley et al., 2017). Recent observations from the SEAC^{4}RS aircraft campaign over the southeast United States in August–September 2013 show much lower NO/NO_2 ratios in the upper troposphere than expected from models (Travis et al., 2016). Here we suggest possible explanations for this discrepancy and discuss the implications for global tropospheric chemistry and for the interpretation of satellite NO_2 data.

The SEAC^{4}RS observations over the southeast United States show NO_x concentrations averaging 0.20 ppb in the upper troposphere above 8 km (0.11 ppb as NO and 0.09 ppb as NO_2), as compared to 0.37 ppb in the boundary layer below 2 km (0.06 ppb NO and 0.31 ppb NO_2), and much lower concentrations (averaging less than 0.07 ppb) in the middle troposphere between 2 and 8 km (Travis et al., 2016). Such a "C-shaped" profile reflects influences from fuel combustion in the boundary layer and lightning in the upper troposphere (Bertram et al., 2007; Hudman et al., 2007; Huntrieser et al., 2002; Pickering et al., 1998). The mean observed daytime NO/NO_2 ratios in SEAC^{4}RS were 0.44 mol mol^{-1} in the boundary layer and 1.4 mol mol^{-1} in the upper troposphere, while the corresponding ratios in the GEOS-Chem chemical transport model sampled along the flight tracks were 0.33 mol mol^{-1} in the boundary layer and 3.3 mol mol^{-1} in the upper troposphere (Travis et al., 2016). The NO/NO_2 ratio in the model increases rapidly with altitude because of the strong temperature dependence of the $NO + O_3$ reaction (Burkholder et al., 2015), but in the observations this increase is much less.

NO_2 measurements in the upper troposphere are prone to positive interferences from inlet decomposition of thermally unstable compounds including HNO_4, $CH_3O_2NO_2$, and other organic nitrates (Bradshaw et al., 1999; Browne et al., 2011; Murphy et al., 2004; Nault et al., 2015; Reed et al., 2016). The Berkeley thermal-dissociation laser-induced fluorescence (TD-LIF) instrument used in SEAC^{4}RS (Day et al., 2002; Thornton et al., 2000; Wooldridge et al., 2010) was specifically designed to minimize and correct for these interferences. Inlet residence time is only 0.23 s, and the NO_2 measurement is corrected for partial thermal dissociation of HNO_4 (0–11%) and $CH_3O_2NO_2$ (0–21%) with a calibration accuracy of 5% (Nault et al., 2015). HNO_4 and $CH_3O_2NO_2$ are independently measured with calibration accuracies of 15%, and the $CH_3O_2NO_2$ measurement has an overall uncertainty of 40%, mainly due to uncertainty in the thermal decomposition rate constant (Nault et al., 2015). An independent NO_2 measurement made by chemiluminescence (Pollack et al., 2010) during SEAC^{4}RS was biased high compared to the TD-LIF measurement, likely due to interferences from $CH_3O_2NO_2$ and HNO_4 (Travis et al., 2016). The Berkeley TD-LIF NO_2 measurements in the upper troposphere have been used in previous work to interpret NO_x chemistry (Nault et al., 2016), lightning NO_x emissions (Nault et al., 2017), and satellite observations of NO_2 columns (Choi et al., 2014; Laughner & Cohen, 2017).

The model-measurement discrepancy in NO/NO_2 partitioning in the upper troposphere has consequences not only for tropospheric chemistry but also for interpreting solar backscatter NO_2 observations from satellites. It is generally assumed that the tropospheric NO_2 column retrieved from satellites is mainly contributed by the boundary layer (Lamsal et al., 2014; Laughner et al., 2016; Martin et al., 2002) and can therefore be related to local NO_x emissions (Martin et al., 2003). However, the NO_2 vertical profiles from SEAC^{4}RS imply a 35–50% contribution of the upper troposphere to the NO_2 tropospheric column observed from satellite (Travis et al., 2016), because sensitivity of backscattered solar radiation to NO_2 increases by a factor of 3 from the surface to the upper troposphere (Martin et al., 2002). Better understanding of this upper tropospheric NO_2 is crucial to the use of satellite data for estimating surface NO_x emissions.

Previous work starting in the 1990s has found varied levels of agreement between NO/NO_2 ratios observed from aircraft and model photochemical equilibrium computed from local conditions. Early work in the lower stratosphere found models to be too high by 20–30% (Cohen et al., 2000; Del Negro et al., 1999; Jaeglé et al., 1994; Sen et al., 1998). Observed NO/NO_2 ratios in the upper troposphere during the PEM-West A campaign over the tropical Pacific were three times lower than model predictions, which was attributed to NO_2 measurement interferences (Crawford et al., 1996), although Davis et al. (1996) also hypothesized a role of halogen chemistry. Bradshaw et al. (1999) found model agreement with observations to within 30% in the upper troposphere over the tropical Pacific using an improved NO_2 instrument that avoided positive interferences through short inlet residence time. More recent model studies of the upper troposphere have again found an overestimate of observed NO/NO_2 ratios (Travis et al., 2016; Williams et al., 2017) and attributed it to underestimate of peroxy radicals converting NO to NO_2.

2. NO-NO_2 Cycling in the Upper Troposphere During SEAC^{4}RS

According to current understanding, the NO/NO_2 ratio in the daytime upper troposphere is determined by rapid chemical cycling through the reactions in Table 1. Here we calculated the mean rates of individual

Table 1
NO-NO_2 Cycling in the Upper Troposphere During SEAC^{4}RS[a]

Reaction	Mean rate (10^6 molecules · cm^{-3} · s^{-1})
Conversion of NO to NO_2	
R1. $NO + O_3 \rightarrow NO_2 + O_2$	3.42 ± 1.04[b–d]
R2. $NO + HO_2 \rightarrow NO_2 + OH$	0.68 ± 0.23[b,d,e]
R3. $NO + CH_3O_2 \rightarrow NO_2 + CH_3O$	0.13 ± 0.04[b,d,e]
R4. $NO + BrO \rightarrow NO_2 + Br$	0.18 ± 0.09[b,d,f]
R5. $NO + IO \rightarrow NO_2 + I$	0.10 ± 0.05[b,d,f]
R6. $NO + ClO \rightarrow NO_2 + Cl$	0.02 ± 0.01[b,d,f]
Total	4.53 ± 1.46
Conversion of NO_2 to NO	
R7. $NO_2 + h\nu \rightarrow NO + O$	8.20 ± 1.74[g]

[a]Main reactions cycling NO and NO_2 in the daytime upper troposphere (8–12 km) over the southeast United States. Mean rates are calculated using JPL kinetic data (Burkholder et al., 2015) applied to SEAC^{4}RS aircraft observations (NO, NO_2, O_3, J_{NO2}, temperature, and pressure) over the southeast United States (94.5–76°W, 30–37°N) in August–September 2013 and with radical concentrations (RO_2, halogens) computed by the GEOS-Chem model along the flight tracks (Sherwen et al., 2016, 2017; Travis et al., 2016). Only reactions with rates above 1×10^4 molecules · cm^{-3} · s^{-1} are listed. Data outside the 9–15 solar time window and in stratospheric air ($[O_3]/[CO] > 1.25$ mol mol^{-1}) have been excluded. Error standard deviations are calculated by propagation of measurement, rate constant, and GEOS-Chem radical concentration errors (precision). [b]Precision of the NO measurement (4%; Ryerson et al., 2000). [c]Precision of the O_3 measurement (3%; Ryerson et al., 1998). [d]Precision of the kinetic rate constants (30% for NO + O_3, 15% for NO + HO_2, NO + CH_3O_2, NO + BrO, and NO + ClO, and 20% for NO + IO; Burkholder et al., 2015). [e]Uncertainty in the GEOS-Chem HO_2 concentration (30%) estimated from the H_2O_2 measurements in SEAC^{4}RS (Crounse et al., 2006). The same relative error is assumed for CH_3O_2. [f]Uncertainty in the GEOS-Chem BrO, IO, and ClO concentrations (50%; Sherwen et al., 2016). [g]Precision of the NO_2 measurement (5%; Nault et al., 2015), the measured UV-A actinic flux (5%; Shetter et al., 2003), and the NO_2 cross section and quantum yield (20%; Sander et al., 2011).

reactions along the SEAC^{4}RS flight tracks over the southeast United States by applying the recommended Jet Propulsion Laboratory (JPL) rate constants (Burkholder et al., 2015) to aircraft measurements of species concentrations (NO, NO_2, and O_3), NO_2 photolysis rate constant (J_{NO2}), temperature, and pressure, together with peroxy and halogen radical concentrations computed by the GEOS-Chem model (Sherwen et al., 2016, 2017; Travis et al., 2016) along the flight tracks. We exclude data outside the 9–15 hr solar time window and in stratospheric air ($[O_3]/[CO] > 1.25$ mol mol^{-1}). We focus on the southeast United States because it accounts for most of the SEAC^{4}RS flights and represents a relatively homogeneous environment.

The calculated rates in Table 1 indicate that conversion of NO to NO_2 in the upper troposphere is mainly by reaction with ozone (75%). Reactions with peroxy radicals contribute 18%, and reactions with halogen radicals (BrO, IO, and ClO) contribute 7%. The total calculated rate of conversion of NO to NO_2 has an aggregated uncertainty of 32% and balances only half of the NO_2 photolysis rate, which has an uncertainty of 21%. This represents a significant discrepancy, such that a model using JPL kinetics would overestimate the NO/NO_2 ratio observed in SEAC^{4}RS.

Figure 1 shows the median observed vertical profiles of the principal variables relevant to NO/NO_2 cycling, along with the corresponding values simulated by the standard GEOS-Chem model along the aircraft flight tracks as in Travis et al. (2016). The NO/NO_2 ratio in the model is over twice that measured in the upper troposphere above 8 km. The bias is systematic over the frequency distribution of the observations (Figure 2) and also extends to stratospherically influenced air ($[O_3]/[CO] > 1.25$ mol mol^{-1}), although SEAC^{4}RS did not sample the actual stratosphere (maximum ozone concentration was 125 ppb). Travis et al. (2016) attributed the model bias in the NO/NO_2 ratio to an underestimate of peroxy radicals, but that underestimate would have to be a factor of 5 in order to close the budget of Table 1. This is incompatible with the SEAC^{4}RS observations for H_2O_2, which is produced by self-reaction of HO_2 and is thus a sensitive proxy of HO_2 concentrations. Observed H_2O_2 concentrations are 30% higher than GEOS-Chem (Figure 1) but would be grossly overestimated if HO_2 concentrations were increased by a factor of 5. Similarly, simulated BrO concentrations would have to be underestimated by a factor of 20 in order to correct the model bias in the NO/NO_2 ratio. This would be grossly inconsistent with observations (Schmidt et al., 2016; Shah et al., 2016; Sherwen et al., 2016).

We see from Figure 1 that there is no systematic model bias in ozone, temperature, or J_{NO2} that would explain an error in the NO/NO_2 ratio. The J_{NO2} observations during SEAC^{4}RS are from spectrally resolved actinic flux measurements (Shetter et al., 2003), converted to photolysis frequencies using the same temperature-dependent JPL spectroscopic data (absorption cross sections and quantum yields) as in GEOS-Chem. These spectroscopic data are tabulated by JPL as a function of wavelength at 220 and 294 K and are interpolated linearly for intermediate temperatures. About 90% of J_{NO2} photolysis is contributed by wavelengths shorter than 398 nm for which the quantum yield is unity. The estimated JPL uncertainty on J_{NO2} is 20% with no temperature dependence (Sander et al., 2011), though laboratory studies show better agreement at surface temperatures (Orphal, 2003; Shetter et al., 2003). Observed photolysis frequencies of other relevant species (O_3, HCHO, H_2O_2, HNO_3, PAN, and CH_3OOH) also agree with GEOS-Chem values to within 1–15% throughout the troposphere.

One possible explanation for the apparent departure of the NO/NO_2 ratio from photochemical equilibrium would be a positive bias in the NO_2 measurement. This could occur if there was an unrecognized labile reservoir of NO_2 (other than HNO_4 or $CH_3O_2NO_2$) decomposing in the instrument inlet, or if the correction for HNO_4 or $CH_3O_2NO_2$ was inadequate. We find that that this missing reservoir (likely organic) would need to

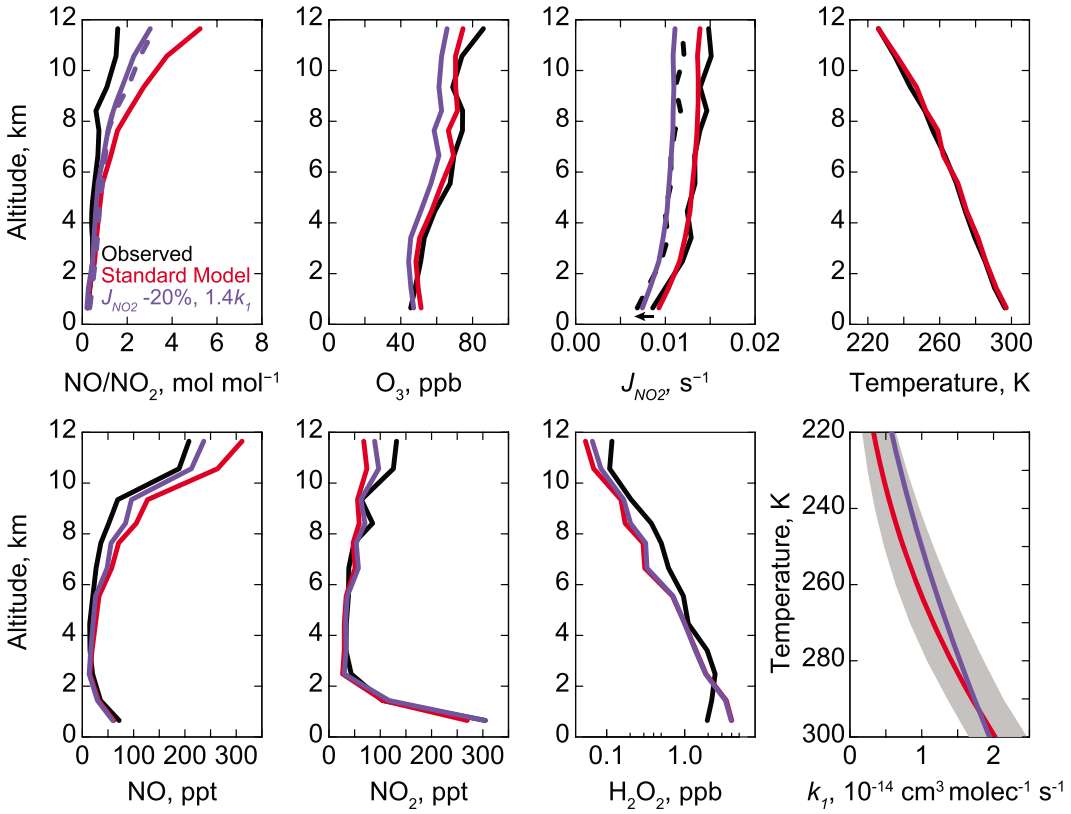

Figure 1. Median vertical tropospheric profiles of the NO/NO_2 concentration ratio and related quantities on SEAC^{4}RS flights (9–15 solar time) during August–September 2013 over the southeast United States (94.5–76°W, 30–37°N). Data from urban plumes ($[NO_2] > 4$ ppb), open fire plumes ($[CH_3CN] > 200$ ppt), and stratospheric air ($[O_3]/[CO] > 1.25$ mol mol^{-1}) are excluded. Observations are compared to GEOS-Chem model results sampled along the flight tracks, for the standard model (Travis et al., 2016), and a sensitivity simulation with reduced J_{NO2} and increased low-temperature NO + O_3 rate constant (see bottom right panel). Observed NO_2 is from the Berkeley TD-LIF measurement (Nault et al., 2015). The bottom right panel shows the k_1 (NO + O_3) rate constant versus temperature from JPL in red with 2-σ uncertainty in gray shading (Burkholder et al., 2015), and sensitivity simulation values resulting in $1.4k_1$ (purple) in the upper troposphere. The J_{NO2} observations apply JPL spectroscopic data to actinic fluxes measured aboard the aircraft and would be reduced similarly to the model if the spectroscopic data are corrected downward (black dashed line). The dashed purple line in the top left panel shows the NO-NO_2-O_3 photochemical equilibrium values $J_{NO2}(h\nu)/k_1(T)[O_3]$ calculated from observed actinic fluxes ($h\nu$), temperature (T), and $[O_3]$ with $1.4k_1$ and the 20% reduction in J_{NO2} applied.

be present at a mean concentration of at least 40 ppt in the upper troposphere in order to fit the model NO/NO_2 ratios, assuming 100% decomposition to NO_2 inside the instrument. For comparison, the correction to the NO_2 measurement from the decomposition of $CH_3O_2NO_2$ was 0–23 ppt (0–21% of NO_2 at temperatures less than 240 K) and 0–20 ppt HNO_4 (0–11%) during SEAC^{4}RS (Nault et al., 2015). The high-flow pump to minimize the influence from these reservoirs malfunctioned in the first five SEAC^{4}RS flights (6–16 August), but these flights were either not over the southeast United States or not in the upper troposphere and are not included in our analysis.

A significant uncertainty in the $CH_3O_2NO_2$ correction is the thermal decomposition rate constant, which has a JPL 1-σ uncertainty of 30%. Considering a cabin air temperature of 300 K, an exterior pressure of 230 hPa, and an inlet residence time of 0.23 s, a 30% increase in the $CH_3O_2NO_2$ thermal decomposition rate would double the corresponding correction to the NO_2 measurement from 0–23 to 0–46 ppt. The effect would be at most 23 ppt at the upper end of the range and is not sufficient to correct the NO/NO_2 ratio. GEOS-Chem underestimates $CH_3O_2NO_2$ in the upper troposphere (18 ± 19 ppt modeled, 124 ± 98 ppt observed), reflecting in part the model underestimate of NO_2 but also suggesting missing organic chemistry. A faster thermal decomposition rate for $CH_3O_2NO_2$ would exacerbate the model underestimate.

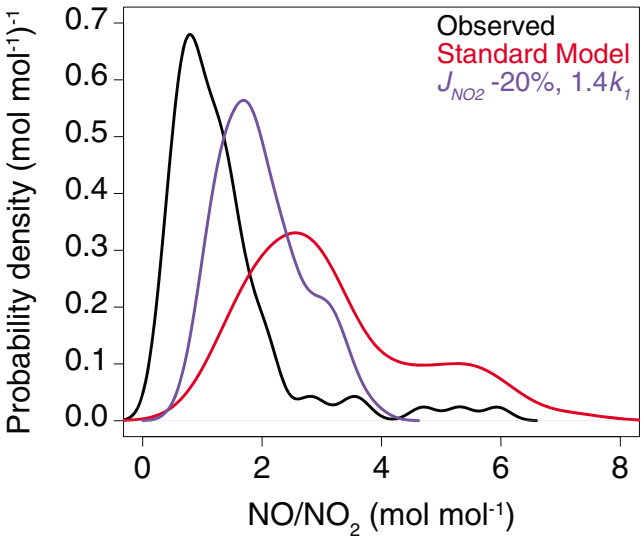

Figure 2. Probability density function of the NO/NO_2 concentration ratio under midday conditions (9–15 local time) in the upper troposphere (8–12 km) during SEAC^{4}RS in August–September 2013. Observations are compared to the standard GEOS-Chem model and the model with J_{NO2} reduced by 20% and the activation energy for the NO + O_3 reaction increased so that k_1 increases on average by 1.4 in the upper troposphere (purple). The same data criteria as stated for Table 1 are applied.

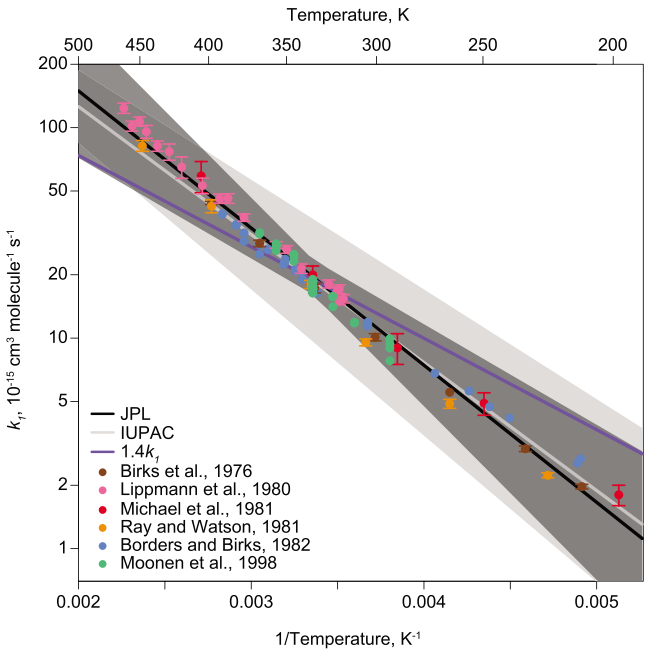

Figure 3. Temperature dependence of the NO + O_3 rate constant k_1. The *y* axis is a log scale, and the *x* axis is an inverse scale (1/*T*), so that an Arrhenius dependence plots as a straight line. Recommended rates from JPL (black; Burkholder et al., 2015) and IUPAC (light gray; Atkinson et al., 2004) are shown as solid lines with the 2-σ uncertainty in shading. Laboratory measurements used in the JPL and IUPAC recommendations are shown in circles with their respective uncertainties. The sensitivity simulation resulting in 1.4k_1 in the upper troposphere (230–250 K) is shown as the purple line.

Previous work has postulated missing organic chemistry in the upper troposphere to explain observations of volatile organic compounds (VOCs) including methanol (Jacob et al., 2005), acetaldehyde (Millet et al., 2010), and glyoxal (Volkamer et al., 2015). Aumont et al. (2005) and Mouchel-Vallon et al. (2013) showed how explicit VOC mechanisms produce a cascade of oxidation products globally that are not tracked in models. Bradshaw et al. (1999) did not need to invoke an unknown NO_x reservoir to reconcile their model with observations over the tropical Pacific, but the southeast United States may be a more propitious environment for VOC oxidation products to be lifted to the upper troposphere by deep convection (Li et al., 2005).

An alternative explanation for the apparent NO/NO_2 departure from photochemical equilibrium would be error in the kinetic data used to compute that equilibrium. Figure 3 shows the International Union of Pure and Applied Chemistry (IUPAC) and JPL recommendations for the temperature dependence of the NO + O_3 rate constant k_1, along with the individual laboratory data that went into these recommendations. IUPAC recommends $k_1 = 2.07 \times 10^{-12}\exp[-1400/T]$ cm^3 · molecule^{-1} · s^{-1} with a 1-σ uncertainty of ±200 K for *E*/*R* and a 1-σ uncertainty of 8% for k_1 at 298 K (Atkinson et al., 2004). JPL recommends $k_1 = 3.00 \times 10^{-12}\exp[-1500/T]$, again with 1-σ uncertainty of ±200 K for *E*/*R* and a 1-σ uncertainty of 10% for k_1 at 298 K. The IUPAC and JPL rate expressions agree to within 4% over the 220–300 K temperature range. For a typical upper tropospheric temperature of 220–240 K the implied 1-σ uncertainty for the JPL rate is 30–40%. However, several studies have suggested a departure from Arrhenius behavior at low temperatures (Figure 3; Birks et al., 1976; Borders & Birks, 1982; Cohen et al., 2000; Michael et al., 1981).

The purple curves in Figures 1 and 2 show the effects in the GEOS-Chem simulation of decreasing J_{NO2} by 20%, and decreasing the activation energy for k_1 by 400 K (2σ) relative to the JPL recommendation so that k_1 increases by a factor 1.4 on average in the upper troposphere, while remaining at the JPL recommended value at 298 K through adjustment of the preexponential factor. The resulting NO/NO_2 ratio in the upper troposphere decreases by 40% from the standard simulation, and Figure 2 shows that the variance in the modeled NO/NO_2 ratio decreases by half, becoming more consistent with the observations. One could match the observations if there was in addition a 15 ppt positive bias in the NO_2 measurement due to an unaccounted labile NO_x reservoir.

We examined whether the variability of the observed NO/NO_2 ratio in the SEAC^{4}RS data set could test the above corrections. For this purpose we used 10-min observations of the ratio along the flight tracks at 8–12 km altitude and correlated them to the local values of the photoequilibrium constant $J_{NO2}(h\nu)/k_1(T)[O_3]$ for $NO\text{-}NO_2\text{-}O_3$ cycling where the UV actinic fluxes ($h\nu$), temperatures (*T*), and [O_3] are taken from the observations. Results are shown in Figure 4. When using JPL values for $J_{NO2}(h\nu)$ and $k_1(T)$, we find a significant ($p < 0.01$) correlation coefficient $r = 0.37$ and a reduced-major-axis regression slope $S = 4.0$. The relatively low correlation coefficient can be attributed to noise and high-frequency variability in the observations. Reducing J_{NO2} by 20% and increasing the low-temperature k_1 as described above improves the slope ($S = 2.3$) while not affecting the correlation coefficient

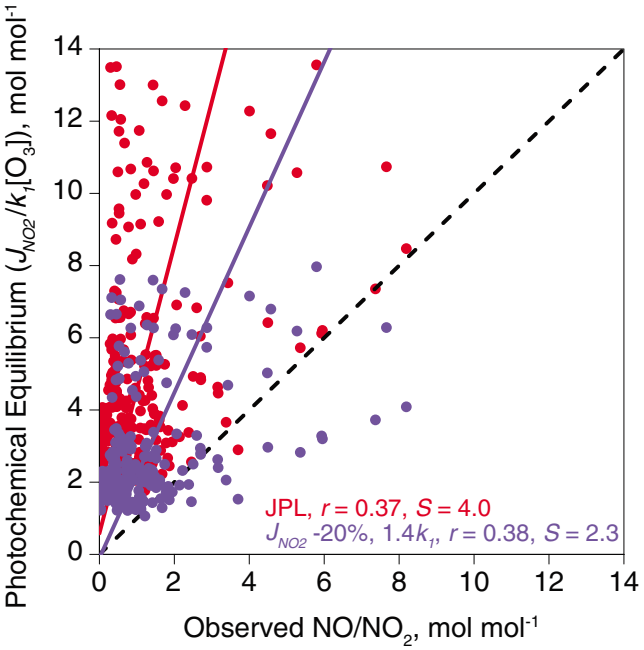

Figure 4. Correlation of the observed NO/NO_2 concentration ratio with the local photochemical equilibrium constant $J_{NO2}(h\nu)/k_1(T)[O_3]$ under midday conditions (9–15 local time) in the upper troposphere (8–12 km) during SEAC^{4}RS in August–September 2013. Data are 10-min averages along the aircraft flight tracks. The photochemical equilibrium constant is calculated from local aircraft measurements of actinic fluxes ($h\nu$), temperature (T), and ozone concentrations. Calculations using the JPL recommendations for $J_{NO2}(h\nu)$ and $k_1(T)$ (in red) are compared to calculations reducing J_{NO2} by 20% and increasing k_1 on average by 1.4 in the upper troposphere (purple). The solid lines show reduced major axis regressions and the 1:1 line is dashed. Correlation coefficients (r) and regression slopes (S) are given inset.

(r = 0.38). Excluding the 15% of the data with $[O_3] < 40$ ppb increases the correlation coefficient to r = 0.48–0.49. Those conditions were associated with particularly high J_{NO2} (reflecting clouds below) and enhancements in CO and HCHO indicative of recent convective influence (Barth et al., 2015; Fried et al., 2016; Snow et al., 2007).

3. Implications

The apparent departure of the NO/NO_2 concentration ratio from photochemical equilibrium in upper tropospheric observations cannot be explained by missing radicals converting NO to NO_2, as proposed in previous work, because the required radical concentrations would be far in excess of observational constraints. It must be due either to an unaccounted labile NO_x reservoir acting as positive interference on the NO_2 measurement or to significant errors in the kinetic data for NO-NO_2-O_3 photochemical cycling at low temperatures. Either of these possibilities have important implications for upper tropospheric chemistry.

The NO_2 measurement specifically excludes interferences from HNO_4 and $CH_3O_2NO_2$, but other labile NO_x reservoirs could potentially be measured as NO_2 following thermal decomposition in the instrument. The presence of such a reservoir at a concentration of 40 ppt, as needed to explain the observed NO/NO_2 ratios, would increase the effective lifetime of NO_x by 20% under the SEAC^{4}RS conditions. More importantly, it would likely imply organic chemistry missing from the models, as also suggested by observations of acetaldehyde and glyoxal in the upper troposphere (Millet et al., 2010; Volkamer et al., 2015) and by the large model underestimate of $CH_3O_2NO_2$ in SEAC^{4}RS.

The rate constants J_{NO2} and k_1 involved in NO-NO_2-O_3 photochemical cycling have relatively small uncertainties in kinetic assessments, and even then are found to be major sources of uncertainty in model simulations of tropospheric oxidants (Bergin et al., 1999; Newsome & Evans, 2017; Ridley et al., 2017; Vuilleumier et al., 1997). When the low-temperature NO + O_3 reaction rate constant ($1.4k_1$) and NO_2 photolysis frequency (J_{NO2} −20%) are adjusted in GEOS-Chem within these uncertainties to improve the simulation of the NO/NO_2 ratio in the SEAC^{4}RS upper tropospheric data, we find that simulated ozone decreases by 7 ppb at 8–12 km altitude. This degrades the previously successful simulation of upper tropospheric ozone in the standard model (Travis et al., 2016); however, that simulation overestimated the NO concentration (Figure 1).

Improved understanding of the contribution of the upper troposphere to the tropospheric NO_2 column also has implications for retrieving and interpreting NO_2 observations from satellites. Spectral fitting of the satellite data in and around the NO_2 absorption bands measures the slant column of NO_2 along the light path. Conversion of this slant column to the actual vertical tropospheric column requires removal of the stratospheric contribution, followed by division by an air mass factor (AMF) dependent on the vertical distribution of tropospheric NO_2 (Martin et al., 2002). The National Aeronautics and Space Administration (NASA) operational retrieval for the OMI satellite instrument (Krotkov et al., 2017) assumes NO_2 vertical profiles from the GMI model (Lamsal et al., 2014) that greatly underestimate NO_2 concentrations in the upper troposphere as observed by SEAC^{4}RS. The mean AMF over the southeast United States in August–September 2013 is 1.28 using vertical distributions from the NASA operational retrieval but 1.67 when using the median observed profile in Figure 1. This implies that the OMI operational retrieval for NO_2 may be 30% too high. Laughner and Cohen (2017) show that inclusion of lightning NO_x in the upper troposphere to match DC3 observations of NO_2 increases the OMI AMF by 34% for summertime, further demonstrating the importance of accurately modeling and measuring NO_x in the upper troposphere for the interpretation of satellite NO_2 measurements.

In conclusion, models significantly overestimate recent observations of the NO/NO_2 ratio in the upper troposphere. This cannot be easily explained by known labile NO_x reservoirs (HNO_4 and $CH_3O_2NO_2$) interfering with the NO_2 measurement. It implies either error in current recommendations for $NO\text{-}NO_2\text{-}O_3$ cycling kinetics or the presence of a missing labile NO_x reservoir, likely organic. Either explanation would have important implications for our understanding of tropospheric oxidants and/or the interpretation of satellite NO_2 measurements.

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