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## RESEARCH ARTICLE

10.1002/2015JD024229

## Key Points:

- Recent BrO observations are interpreted using a new GEOS-Chem coupled Br-Cl simulation
- Multiphase oxidation of Br- by ozone is critical for maintaining the high observed levels of BrO
- Br and Cl lower the global burden of O<sub>3</sub> by 14% by increasing NO<sub>x</sub> loss and shortening the O<sub>3</sub> lifetime

## Supporting Information:

- Supporting Information SI

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## Modeling the observed tropospheric BrO background: Importance of multiphase chemistry and implications for ozone, OH, and mercury

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**Abstract** Aircraft and satellite observations indicate the presence of ppt (ppt ≡ pmol/mol) levels of BrO in the free troposphere with important implications for the tropospheric budgets of ozone, OH, and mercury. We can reproduce these observations with the GEOS-Chem global tropospheric chemistry model by including a broader consideration of multiphase halogen (Br-Cl) chemistry than has been done in the past. Important reactions for regenerating BrO from its nonradical reservoirs include HOBr + Br<sup>-</sup>/Cl<sup>-</sup> in both aerosols and clouds, and oxidation of Br<sup>-</sup> by ClNO<sub>2</sub> and ozone. Most tropospheric BrO in the model is in the free troposphere, consistent with observations and originates mainly from the photolysis and oxidation of ocean-emitted CHBr<sub>3</sub>. Stratospheric input is also important in the upper troposphere. Including production of gas phase inorganic bromine from debromination of acidified sea salt aerosol increases free tropospheric Br<sub>y</sub> by about 30%. We find HOBr to be the dominant gas-phase reservoir of inorganic bromine. Halogen (Br-Cl) radical chemistry as implemented here in GEOS-Chem drives 14% and 11% decreases in the global burdens of tropospheric ozone and OH, respectively, a 16% increase in the atmospheric lifetime of methane, and an atmospheric lifetime of 6 months for elemental mercury. The dominant mechanism for the Br-Cl driven tropospheric ozone decrease is oxidation of NO<sub>x</sub> by formation and hydrolysis of BrNO<sub>3</sub> and ClNO<sub>3</sub>.

## 1. Introduction

Atmospheric bromine radicals (BrO<sub>x</sub> ≡ Br + BrO) destroy ozone, perturb HO<sub>x</sub> (OH + HO<sub>2</sub>) and NO<sub>x</sub> (NO + NO<sub>2</sub>) radical budgets, and provide a sink for mercury [von Glasow et al., 2004; Parrella et al., 2012; Simpson et al., 2015]. The importance of bromine radical chemistry is well established in the stratosphere [Montzka et al., 2010] and in the Arctic boundary layer in spring [Barrie et al., 1988; Simpson et al., 2007]. There is increasing evidence for a tropospheric background BrO concentration of the order of 1 ppt (ppt ≡ pmol/mol) in the daytime [Platt and Hönninger, 2003; Sinnhuber et al., 2005; Prados-Roman et al., 2011; Theys et al., 2011; Volkamer et al., 2015; Wang et al., 2015]. Such a background would provide a major sink for tropospheric ozone and also deplete OH, the main tropospheric oxidant [von Glasow et al., 2004; Yang et al., 2005; Saiz-Lopez et al., 2012; Parrella et al., 2012; Wang et al., 2015]. It would also imply atomic Br concentrations sufficiently high to provide the main oxidant for gaseous elemental mercury and thus drive the patterns of mercury deposition to ecosystems [Holmes et al., 2010; Wang et al., 2015; Gratz et al., 2015; Coburn et al., 2016]. Here we use a global 3-D chemical transport model (GEOS-Chem CTM) to interpret recent observations of tropospheric BrO and its reservoirs in terms of our understanding of tropospheric bromine chemistry and its implications.

The main sources of reactive inorganic bromine (Br<sub>y</sub>) in the troposphere are photochemical decomposition of organobromines (CHBr<sub>3</sub>, CH<sub>2</sub>Br<sub>2</sub>, CH<sub>3</sub>Br), release of bromine from sea salt aerosol (SSA), and transport from

the stratosphere where  $\text{Br}_y$  originates from photochemical decomposition of organobromines and halons [Yang *et al.*, 2005; Liang *et al.*, 2014].  $\text{CH}_2\text{Br}_2$  and  $\text{CHBr}_3$  are of marine biogenic origin, while  $\text{CH}_3\text{Br}$  has both natural and anthropogenic sources [Montzka *et al.*, 2010]. Cycling of  $\text{Br}_y$  takes place between  $\text{BrO}_x$  radicals and nonradical reservoirs (including  $\text{HBr}$ ,  $\text{HOBr}$ ,  $\text{BrNO}_2$ ,  $\text{BrNO}_3$ ,  $\text{Br}_2$ , and  $\text{BrCl}$ ).  $\text{Br}_y$  is eventually removed from the troposphere by dry and wet deposition.

A number of global model studies have pointed out the potential importance of  $\text{BrO}_x$  radicals for global tropospheric chemistry [von Glasow *et al.*, 2004; Yang *et al.*, 2005; Warwick *et al.*, 2006; Breider *et al.*, 2010; Hossaini *et al.*, 2010; Parrella *et al.*, 2012; Ordóñez *et al.*, 2012; Saiz-Lopez *et al.*, 2012; Long *et al.*, 2014; Fernandez *et al.*, 2014; Sherwen *et al.*, 2016]. Confidence in these models has been limited by the paucity of observational constraints on  $\text{BrO}$  and other  $\text{Br}_y$  species. The previous GEOS-Chem study by Parrella *et al.* [2012] underestimated GOME-2 satellite observations of tropospheric  $\text{BrO}$  columns by  $\sim 30\%$ . Model predictions for  $\text{BrO}$  in the tropical upper troposphere [Yang *et al.*, 2005; Parrella *et al.*, 2012; Long *et al.*, 2014; Fernandez *et al.*, 2014] are much smaller than recent observations [Volkamer *et al.*, 2015; Wang *et al.*, 2015]. Here we use new aircraft observations of  $\text{BrO}$ , together with  $\text{BrO}$  observations from satellites, to better constrain tropospheric bromine chemistry in models. We propose in particular that the coupling of bromine reservoirs ( $\text{HBr}/\text{Br}^-$  and  $\text{HOBr}$ ) to chlorine and ozone through multiphase chemistry is more important than previously recognized for sustaining the high observed levels of  $\text{BrO}$  and  $\text{Br}_y$  in the troposphere.

## 2. Model Description

We use the GEOS-Chem global CTM (v9-02; <http://www.geos-chem.org>) including a detailed ozone- $\text{NO}_x$ -VOC-aerosol- $\text{Br-Cl}$  tropospheric chemistry mechanism. The model is driven by GEOS-5 assimilated meteorological data from the NASA Global Modeling and Assimilation Office with  $1/2^\circ \times 2/3^\circ$  horizontal resolution and 47 vertical layers extending from the surface up to 80 km. The horizontal resolution is degraded here to  $4^\circ \times 5^\circ$  for input to GEOS-Chem. The year 2007 is chosen as a reference and all simulations are spun-up over 1 year (2006) for initialization.

The simulation updates the tropospheric bromine mechanism originally described by Parrella *et al.* [2012], to include a more extensive multiphase chemistry mechanism as described below. We have also added a simulation of tropospheric chlorine radical chemistry coupled to that of bromine as providing an important pathway for recycling of bromine radicals. For the purpose of this paper we define  $\text{Br}_y$  and  $\text{Cl}_y$  as the reactive inorganic halogens, excluding halide contained in sea salt that has not yet been activated to produce gas-phase inorganic halogen species.

The sources of reactive tropospheric inorganic halogens are listed in Table 1. Production from organohalogen by oxidation and photolysis is well established. Our simulation of organobromines is as in Parrella *et al.* [2012], who showed good agreement with vertical profiles observed from aircraft campaigns. We confirm this agreement in model comparisons to aircraft observations of organobromines from the CARIBIC project covering the upper troposphere [Wisher *et al.*, 2014] (see also <http://www.caribic-atmospheric.com>), the TORERO campaign covering the Eastern Pacific, and HIPPO campaign across the Pacific [Wofsy *et al.*, 2012a, 2012b], see Figure S1 in the supporting information (SI).  $\text{CH}_3\text{Br} + \text{OH}$  kinetics are updated according to Nilsson *et al.* [2013], increasing the  $\text{Br}_y$  source contribution by tropospheric  $\text{CH}_3\text{Br}$  from  $56 \text{ Gg Br a}^{-1}$  in Parrella *et al.* [2012] to  $91 \text{ Gg Br a}^{-1}$  here. This remains small compared to the dominant organobromine source from  $\text{CHBr}_3$  ( $404 \text{ Gg Br a}^{-1}$ ). Stratospheric  $\text{Br}_y$  is treated as a boundary condition above the GEOS-5 tropopause using archived  $\text{Br}_y$  fields from the GEOSCCM global 3-D simulation by Liang *et al.* [2010] that are consistent with balloon-based stratospheric  $\text{BrO}$  observations [Liang *et al.*, 2014].

Additional generation of bromine radicals by oxidation of bromide in SSA is uncertain and highly variable. Observations in the marine boundary layer (MBL) generally show  $\text{BrO}$  below the detection limit of  $\leq 1$  ppt [Leser *et al.*, 2003; Gomez Martin *et al.*, 2013; Volkamer *et al.*, 2015; Wang *et al.*, 2015], but some studies show detectable higher values [Saiz-Lopez *et al.*, 2004; Read *et al.*, 2008; Martin *et al.*, 2009]. The underlying mechanisms are poorly understood and could be sensitive to a number of factors including aerosol acidity and the presence of dissolved organic matter [Fickert *et al.*, 1999; Liang and Singer, 2003]. We perform two separate simulations (A and B): Our simulation A does not include dehalogenation of SSA. Our simulation B includes dehalogenation of acidic SSA as a source of  $\text{Br}_y$  and  $\text{Cl}_y$ . The GEOS-Chem SSA simulation is as described by Jaegle *et al.* [2011] and sea salt bromide ( $\text{Br}_{\text{SSA}}^-$ ) is emitted as part of SSA with a ratio of  $2.11 \times 10^{-3} \text{ kg Br (kg dry SSA)}^{-1}$  [Lewis and Schwartz, 2004]. The transport and deposition of  $\text{Br}_{\text{SSA}}^-$  follows that of the parent

**Table 1.** Global Sources of Reactive Inorganic Halogens to the Troposphere

	Br <sub>y</sub> /Gg Br a <sup>-1</sup>	Cl <sub>y</sub> /Gg Cl a <sup>-1</sup>
Sea salt <sup>a</sup>	(1620)	(6050)
Organohalogens <sup>b</sup>		
CH <sub>3</sub> X	91	2350
CH <sub>2</sub> X <sub>2</sub>	55	483
CHX <sub>3</sub>	404	262
Stratosphere	49	407

<sup>a</sup>Release of inorganic bromine and chlorine from sea salt as included in simulation B (see text).

<sup>b</sup>X ≡ Br or Cl.

sea salt aerosol. Release of Br<sub>2</sub>, BrCl, and HOBr from acidified SSA follows the general multiphase chemistry mechanism described below. Uptake of gas-phase HBr by SSA provides an additional source of Br<sub>SSA</sub><sup>-</sup>.

Removal of Br<sub>y</sub> and Cl<sub>y</sub> takes place by wet and dry deposition. Wet deposition of gases and aerosols follows the schemes described by *Amos et al.* [2012] and *Wang et al.* [2011], respectively. Dry deposition is computed with the resistance-in-series scheme

of *Wesely* [1989] as implemented by *Wang et al.* [1998]. Deposition of gases depends on the Henry's law and acid dissociation constants.

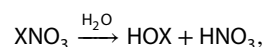
### 2.1. Tropospheric Chlorine Simulation

The chlorine simulation is based on the GEOS-Chem UCX chlorine mechanism for the stratosphere [*Eastham et al.*, 2014] extended here to the troposphere. Sources of reactive inorganic tropospheric chlorine (Cl<sub>y</sub>) include oxidation of organochlorines, release from sea salt (in simulation B), and transport from the stratosphere (Table 1). We add 14 gas phase chlorine species to the GEOS-Chem tropospheric mechanism: BrCl, Cl<sub>2</sub>, Cl, ClO, HCl, HOCl, ClNO<sub>2</sub>, ClNO<sub>3</sub>, ClOO, OCIO, Cl<sub>2</sub>O<sub>2</sub>, CH<sub>3</sub>Cl, CH<sub>2</sub>Cl<sub>2</sub>, and CHCl<sub>3</sub>.

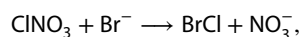
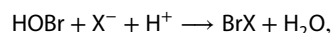
The organochlorines release atomic Cl when oxidized by OH and Cl. Their sources are represented implicitly by imposing fixed surface mixing ratios of 550 ppt CH<sub>3</sub>Cl, 20 ppt CH<sub>2</sub>Cl<sub>2</sub>, and 7 ppt CHCl<sub>3</sub> based on 2007 NOAA and AGAGE network data [*Montzka et al.*, 2010]. The model lifetimes of these organochlorines against oxidation are 480, 158, and 152 days, respectively, which implies that CH<sub>3</sub>Cl is the dominant source of inorganic chlorine in the standard simulation (See Table 1). Chlorine is released from SSA in simulation B through the multiphase chemistry mechanism described below. We do not consider acid displacement of HCl from sea salt nor chloride activation by N<sub>2</sub>O<sub>5</sub>. The latter mechanism can drive large chlorine release in polluted continental regions [*Thornton et al.*, 2010], but we expect it to be of little importance globally. Stratospheric chlorine is simulated following *Murray et al.* [2012] by using archived monthly mean production and loss rate constants for individual chlorine species from the Global Modeling Initiative (GMI) stratospheric model.

### 2.2. Halogen Multiphase Chemistry

Our halogen multiphase chemistry mechanism (halogen X ≡ Br or Cl) is based on recommendations by the International Union of Pure and Applied Chemistry (IUPAC) [*Ammann et al.*, 2013] and takes place in liquid and ice clouds, sea salt aerosol, and sulfate aerosol (the implementation of sulfate aerosols in GEOS-Chem is described by *Park et al.* [2004] and *Pye et al.* [2009]). The mechanism includes hydrolysis of halonitrates,



and oxidation of halides by HOBr, ClNO<sub>3</sub>, and O<sub>3</sub>,



The rate of reaction depends on the concentration of gas-phase reactant, [A(g)], and condensed phase surface area concentration, [S], following the parameterization from *Ammann et al.* [2013],

$$\frac{d[A(g)]}{dt} = -\gamma \frac{c}{4} [S][A(g)], \quad (1)$$

**Table 2.** Reactive Uptake Coefficients for Halogen Multiphase Hydrolysis<sup>a</sup>

Reaction	Reactive Uptake Coefficient
$\text{BrNO}_3 \xrightarrow{\text{H}_2\text{O}} \text{HOBr} + \text{HNO}_3$	$\gamma' = (1/\Gamma_b + 1/\alpha_b)^{-1}$ $\Gamma_b = 0.03$ ; $\alpha_b = 0.063$
$\text{ClNO}_3 \xrightarrow{\text{H}_2\text{O}} \text{HOCl} + \text{HNO}_3$	$\gamma' = (1/\Gamma_b + 1/\alpha_b)^{-1}$ $\Gamma_b = 0.03$ ; $\alpha_b = 0.11$

<sup>a</sup>Parameters from Ammann *et al.* [2013] unless otherwise stated.  $\gamma'$  is defined implicitly from equation (2) and is equal to the reactive uptake coefficient ( $\gamma$ ) when gas phase diffusion is not limiting.  $\alpha_b$  and  $\Gamma_b$  are the mass accommodation coefficient and the bulk reaction coefficient, respectively. The previous GEOS-Chem study [Parrella *et al.*, 2012] included  $\text{BrNO}_3$  hydrolysis with  $\gamma' = 0.3$  for sea salt and liquid cloud droplets and 0.8 for sulfate aerosol. Our  $\Gamma_b$  is estimated from observed  $\gamma'$  reported by Deiber *et al.* [2004].

where  $\gamma$  is the reactive uptake coefficient and  $c$  is the mean thermal velocity of A. The reactive uptake coefficient as defined here includes all mass transfer limitations. It is calculated following a resistor-in-series expression,

$$\frac{1}{\gamma} = \frac{1}{\gamma_d} + \frac{1}{\gamma'} \quad (2)$$

where  $1/\gamma_d$  is the resistance to gas phase diffusion that depends on the particle radius ( $r$ ) and the gas diffusion coefficient of A in air ( $D_{A,g}$ ),

$$\gamma_d = \frac{4 D_{A,g}}{c r}. \quad (3)$$

**Table 3.** Reactive Uptake Coefficients for Multiphase Oxidation of Halide Ions<sup>a</sup>

Reaction	Reactive Uptake Coefficient
$\text{HOBr} + \text{Br}^- + \text{H}^+ \rightarrow \text{Br}_2 + \text{H}_2\text{O}$	$\gamma' = (1/\Gamma_b + 1/\alpha_b)^{-1}$ $\Gamma_b = 4 H_{\text{HOBr}} RT l_r k_{\text{bulk}} [\text{Br}^-] [\text{H}^+] f(r, l_r) / c$ $l_r = \sqrt{D_I / (k_{\text{bulk}} [\text{Br}^-] [\text{H}^+])}$ ; $\alpha_b = 0.6$ $k_{\text{bulk}} = 1.6 \times 10^{10} \text{ M}^{-2} \text{ s}^{-1}$ ; $D_I = 1.4 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$
$\text{HOBr} + \text{Cl}^- + \text{H}^+ \rightarrow \text{BrCl} + \text{H}_2\text{O}$	$\gamma' = (1/\Gamma_b + 1/\alpha_b)^{-1}$ $\Gamma_b = 4 H_{\text{HOBr}} RT l_r k_{\text{bulk}} [\text{Cl}^-] [\text{H}^+] f(r, l_r) / c$ $l_r = \sqrt{D_I / (k_{\text{bulk}} [\text{Cl}^-] [\text{H}^+])}$ ; $\alpha_b = 0.6$ $k_{\text{bulk}} = 5.9 \times 10^9 \text{ M}^{-2} \text{ s}^{-1}$ ; $D_I = 1.4 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$
$\text{ClNO}_3 + \text{Br}^- \rightarrow \text{BrCl} + \text{NO}_3^-$	$\gamma' = (1/\Gamma_b + 1/\alpha_b)^{-1}$ $\Gamma_b = 4 W RT \sqrt{[\text{Br}^-] D_I} / c$ ; $\alpha_b = 0.108$ $D_I = 5.0 \times 10^{-6} \text{ cm}^2 \text{ s}^{-1}$ ; $W = 10^6 \sqrt{\text{M}} \text{ s bar}^{-1}$
$\text{O}_3 + \text{Br}^- \xrightarrow{\text{H}_2\text{O}} \text{HOBr} + \text{OH}^- + \text{O}_2$	$\gamma' = \Gamma_b + \Gamma_s$ $\Gamma_b = 4 H_{\text{O}_3} RT l_r k_{\text{bulk}} [\text{Br}^-] f(r, l_r) / c$ $l_r = \sqrt{D_I / (k_{\text{bulk}} [\text{Br}^-])}$ ; $D_I = 8.9 \times 10^{-6} \text{ cm}^2 \text{ s}^{-1}$ $k_{\text{bulk}} = 6.3 \times 10^8 \text{ M}^{-1} \text{ s}^{-1} \exp(-4450 \text{ K} / T)$ $\Gamma_s = (4 k_s [\text{Br}^- (\text{surf})] K_{\text{LangC}} N_{\text{max}}) / (c_{\text{avg}} (1 + K_{\text{LangC}} [\text{O}_3 (\text{g})]))$ $[\text{Br}^- (\text{surf})] = \min(3.41 \times 10^{14} \text{ cm}^{-2} \text{ M}^{-1} [\text{Br}^-], N_{\text{max}})$ $k_s = 10^{-16} \text{ cm}^2 \text{ s}^{-1}$ ; $K_{\text{LangC}} = 10^{-13} \text{ cm}^3$ ; $N_{\text{max}} = 3.0 \times 10^{14} \text{ cm}^{-2}$

<sup>a</sup>Parameters from Ammann *et al.* [2013] unless otherwise stated. The previous GEOS-Chem study [Parrella *et al.*, 2012] included multiphase  $\text{HOBr} + \text{HBr}$  with  $\gamma' = 0.2$  for sea salt and sulfate aerosol, and 0.1 for ice cloud droplets.  $k_{\text{bulk}}$  for  $\text{HOBr} + \text{Br}^-$  is taken from Beckwith *et al.* [1996]. For sea salt aerosol  $\text{HOBr} + \text{Cl}^-$  is assumed to be limited by mass accommodation or gas phase diffusion kinetics.

The expression  $1/\gamma'$  is the total condensed-phase resistance to subsequent mass accommodation, mass transfer, and first-order reaction. It depends on a number of factors including particle radius, particle halide concentration, pH, and temperature, as described in Tables 2 and 3. In addition, we include uptake of HBr on SSA to form  $\text{Br}_{\text{SSA}}^-$  with  $\gamma' = 1.3 \times 10^{-8} \exp(4290 \text{ K} / T)$ .

Generation of volatile dihalogens from the aqueous-phase reaction of HOBr with bromide and chloride (R1) is pH dependent and requires acidic conditions [Fickert *et al.*, 1999]. Cloudwater pH (typically in range of 4 to 6) is calculated locally in GEOS-Chem following Alexander *et al.* [2012]. Sulfate aerosol is assumed to have a pH of 0 following the observations by Froyd *et al.* [2009] who found sulfate aerosol in the free troposphere to be strongly acidic (pH < 1). Sulfate aerosol in the MBL is also prevalently acidic [Paulot *et al.*, 2015]. Sea salt aerosol is emitted alkaline, but the alkalinity can be titrated in GEOS-Chem by uptake of  $\text{HNO}_3$ ,  $\text{SO}_2$ , and  $\text{H}_2\text{SO}_4$  [Alexander *et al.*, 2005]. Sea salt aerosol with retained alkalinity is assumed to have pH = 8 (similar to sea water), while sea salt aerosol with no remaining alkalinity is assumed to have a pH of 5. In simulation B sea salt dehalogenation is enabled for acidic SSA.

The cloud droplet and sulfate aerosol halide concentration,  $[\text{X}^-]$ , is modeled assuming local equilibrium between gas phase HX and condensed phase  $\text{X}^-$ . We model the surface area concentration of liquid and ice clouds using local liquid and ice water contents (LWC and IWC) from the GEOS-5 meteorological data and assuming effective radii of 10  $\mu\text{m}$  and 75  $\mu\text{m}$ , respectively. Studies of ice crystals have shown the existence of an unfrozen overlayer coating the droplets, in which halogens accumulate as they are expelled from the ice lattice [Bogdan *et al.*, 2006, 2010]. The thickness of this layer is typically  $10^{-3}$  to  $10^{-2}$  of the ice crystal diameter; in the present simulation we assume a constant value of  $10^{-2}$ . Ice cloud multiphase chemistry is assumed to be confined to this layer.

### 3. Global Distribution of Tropospheric Bromine

Figure 1 (top) shows the GEOS-Chem simulated annual budget of tropospheric inorganic bromine, for the simulation not including SSA dehalogenation (simulation A). The total  $\text{Br}_y$  concentration is 7% less than the previous GEOS-Chem simulation by Parrella *et al.* [2012]. That study included a sea salt debromination source based on observed bromide depletion factors [Sander *et al.*, 2003], with a corresponding global  $\text{Br}_y$  source from sea salt debromination of 1420  $\text{Gg Br a}^{-1}$  that is similar to the source of 1620  $\text{Gg Br a}^{-1}$  in our simulation B that includes dehalogenation of acidic SSA (Table 1). However, the geographic distribution of SSA debromination is different; Parrella *et al.* [2012] found debromination over the Southern Ocean to dominate due to high-emission sea salt in this region, whereas we find less debromination in this region as acid input is often insufficient to compensate for the high-alkalinity flux. Our simulation including SSA debromination show highest levels of BrO in the MBL over the tropical and North Atlantic, where SSA alkalinity is generally fully titrated. We find a fourfold increase in  $\text{Br}_y$  in the MBL when SSA debromination is included. The effect of SSA debromination on the free troposphere is smaller, increasing  $\text{Br}_y$  by about 30% (Figure 1). Approximately 1/4 of free tropospheric  $\text{Br}_y$  in simulation B can be attributed to SSA debromination, the dominant source of  $\text{Br}_y$  in the free troposphere is photochemical oxidation of  $\text{CHBr}_3$  and input from the stratosphere. HOBr is the dominant daytime  $\text{Br}_y$  species. BrCl and  $\text{BrNO}_3$  are the dominant nighttime species. Overall, total  $\text{Br}_y$  show almost no diurnal variability.

The mean tropospheric BrO concentration in our simulation A is 0.48 ppt (0.96 ppt in daytime), 50% higher than Parrella *et al.* [2012] and more consistent with the observed range as shown below. The higher BrO/ $\text{Br}_y$  ratio reflects the more efficient multiphase HBr recycling. HBr accounts for only 6% of  $\text{Br}_y$  in our simulation, as compared to 34% in Parrella *et al.* [2012]. Cycling of HBr in our simulation is faster as we include bromide oxidation by HOBr in liquid cloud droplets and by  $\text{ClNO}_3$  and  $\text{O}_3$ , not considered in Parrella *et al.* [2012]. We find that  $\text{BrNO}_3$  is a more important  $\text{Br}_y$  reservoir as aqueous-phase  $\text{BrNO}_3$  hydrolysis in our simulation ( $\gamma' = 0.02$ ) is considerably slower than in Parrella *et al.* [2012] ( $\gamma' = 0.3$  for clouds and sea salt and  $\gamma' = 0.8$  for sulfate aerosol).

The model study of Long *et al.* [2014] found  $\text{Br}_2$  and BrCl to be the dominant tropospheric  $\text{Br}_y$  species globally, and the study of Fernandez *et al.* [2014] found HBr to dominate. We find HOBr to be the dominant  $\text{Br}_y$  reservoir globally (Figure 1), except in the upper troposphere where HOBr formation by the  $\text{BrO} + \text{HO}_2$  reaction is suppressed by low levels of  $\text{HO}_2$  (See Figures S4 and S5). BrCl is the dominant nighttime  $\text{Br}_y$  reservoir, produced by the  $\text{HOBr} + \text{Cl}^-$  multiphase reaction, but  $\text{BrNO}_3$  becomes increasingly important in the upper

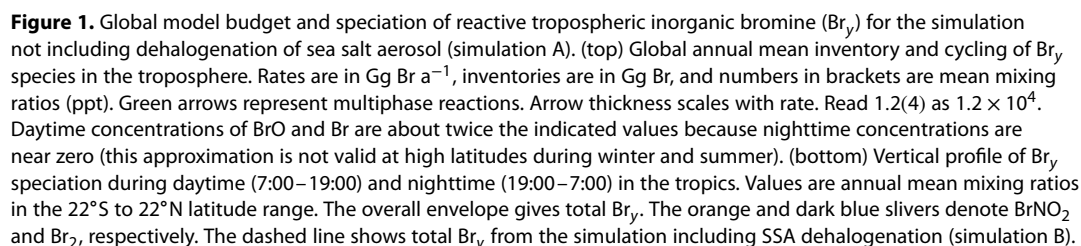
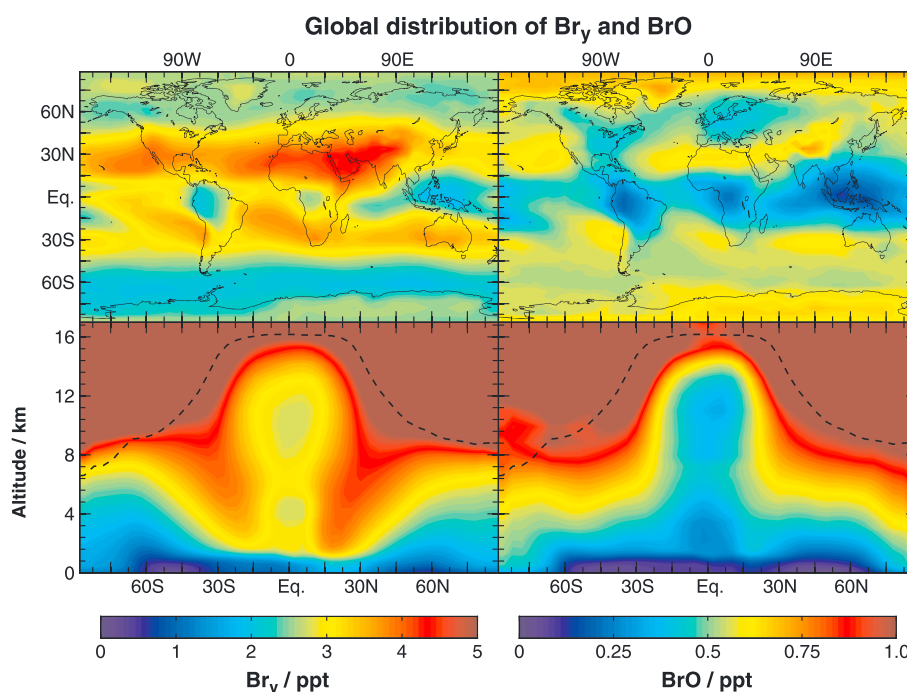


Figure 2 shows the global mean distributions of tropospheric  $\text{Br}_y$  and  $\text{BrO}$  (simulation A).  $\text{Br}_y$  is largest in the subsiding subtropics due to the high-altitude source from organobromine photochemistry and the lack of wet deposition. Input from the stratosphere is also a significant source in that region, accounting for most of  $\text{Br}_y$  above 7 km.  $\text{BrO}$  is relatively high in the subtropics and in the upper troposphere, reflecting the distribution of  $\text{Br}_y$  but also shows different patterns driven by the abundance of  $\text{HO}_2$ . Thus, the high levels of  $\text{BrO}$  in polar regions reflect low  $\text{HO}_2$  levels pushing the  $\text{BrO}/\text{HOBr}$  equilibrium toward  $\text{BrO}$  (Figure S3). Photochemical equilibrium maintains  $[\text{BrO}]/[\text{Br}] \gg 1$  throughout the troposphere (Figure 1) except in the tropical tropopause layer above 15 km where the  $[\text{BrO}]/[\text{Br}]$  ratio decreases to about unity (See SI Text S1 and Figure S4). This decrease is driven by low temperatures and low concentrations of ozone in the tropical upper troposphere suppressing the  $\text{Br} + \text{O}_3 \rightarrow \text{BrO} + \text{O}_2$  reaction. This enhancement of  $\text{Br}$  in the tropical upper troposphere was previously identified by *Holmes et al.* [2006] for its importance in  $\text{Hg}(0)$  oxidation and is a consistent feature of models [*Fernandez et al.*, 2014].



**Figure 2.** Simulated annual mean global distributions of tropospheric  $\text{Br}_y$  and  $\text{BrO}$ . Multiply  $\text{BrO}$  concentrations by a factor of 2 for daytime values (this approximation is not valid at high latitudes during winter and summer). (top row) Mean tropospheric mixing ratios. (bottom row) Zonal mean mixing ratios as a function of altitude and latitude. The tropopause is shown as dashed line. Results are for the simulation not including dehalogenation of SSA (simulation A).

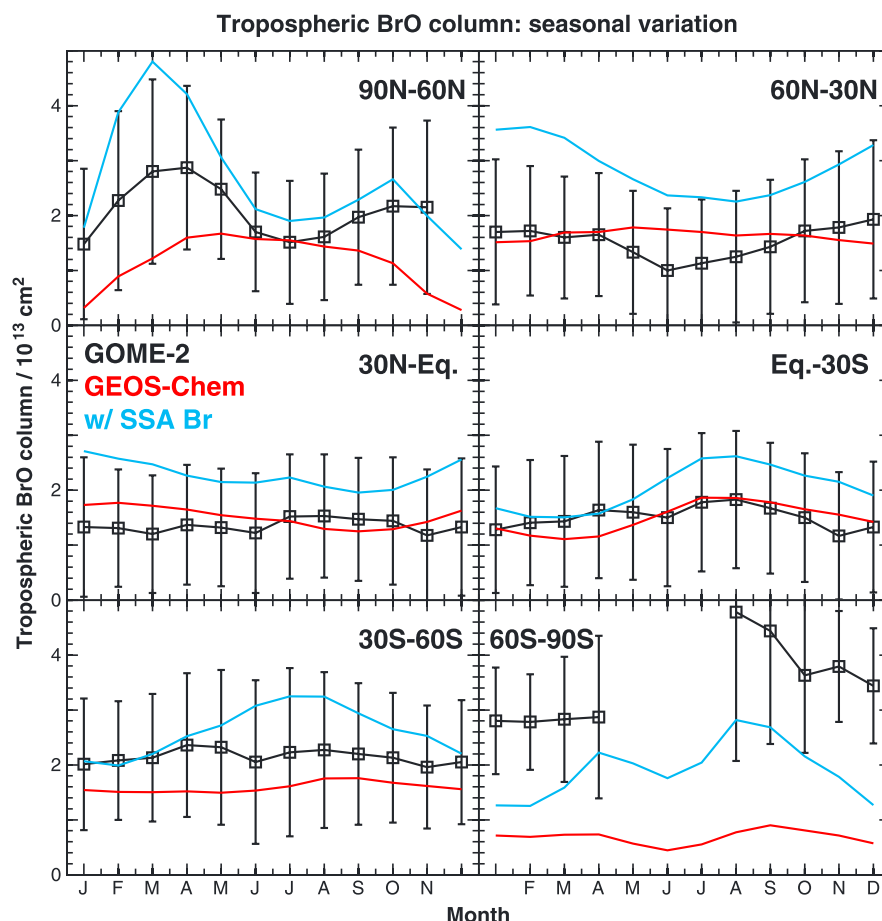
The global mean distribution  $\text{Br}_y$  and  $\text{BrO}$  from the simulation B (that includes dehalogenation of acidic SSA) is shown in Figure S8. The increase in  $\text{Br}_y$  and  $\text{BrO}$  from including SSA dehalogenation is particularly strong in the Northern Hemisphere MBL. Simulated SSA in this region tend to be more depleted in bromide (Figure S9). SSA over Southern Ocean tends to retain its alkalinity preventing dehalogenation.

#### 4. Comparison to $\text{BrO}$ and $\text{HOBr}$ Observations

Figure 3 compares the simulated seasonal distribution of tropospheric  $\text{BrO}$  column concentrations in different latitude bands to GOME-2 satellite observations of *Theys et al.* [2011]. Both observations and model values are for 2007. Simulation A is consistent with observations in the tropics and at midlatitudes but too low in polar regions. The previous GEOS-Chem model study of *Parrella et al.* [2012] found a  $\sim 50\%$  low bias in the tropics. *Theys et al.* [2011] inferred the vertical profile of tropospheric  $\text{BrO}$  in the tropics ( $30^\circ\text{S}$ – $30^\circ\text{N}$ ) using a cloud-slicing technique and found that over 75% of the tropospheric  $\text{BrO}$  column is above 2 km. Both simulations (A and B) are consistent with this finding (Figure 1). Observations in the extratropical Northern Hemisphere show a summer minimum that was well simulated by *Parrella et al.* [2012] as due to seasonality in the SSA source. Our simulation including SSA dehalogenation captures this observed seasonal variation but is overall too high.

Figure 4 compares model results to the total  $\text{BrO}$  column observed by the OMI satellite instrument (including the stratosphere) as a function of latitude. The simulated total  $\text{BrO}$  column includes GEOS-Chem tropospheric  $\text{BrO}$  plus GEOSCCM stratospheric  $\text{BrO}$  from *Liang et al.* [2010], which serves as upper boundary condition for GEOS-Chem. We find close agreement with observations. Tropospheric  $\text{BrO}$  contributes between 25% and 55% of the total  $\text{BrO}$  column. Both GEOS-Chem and observations show lowest values in regions of tropical upwelling, where tropospheric  $\text{Br}_y$  is efficiently scavenged, although this is exaggerated in GEOS-Chem.

The first vertically resolved detection of tropospheric  $\text{BrO}$  in the tropical troposphere was demonstrated by the University of Colorado Airborne Multi-Axis Differential Optical Absorption Spectroscopy (CU AMAX-DOAS) during the TORERO aircraft and ship campaign over the East Pacific (January–February 2012) [*Volkamer et al.*, 2015]. Model results sampled along all 17 flight tracks are compared to observations in Figure 5, separating tropical from subtropical flights. No  $\text{BrO}$  was detected in the MBL over the tropical Eastern Pacific. The upper

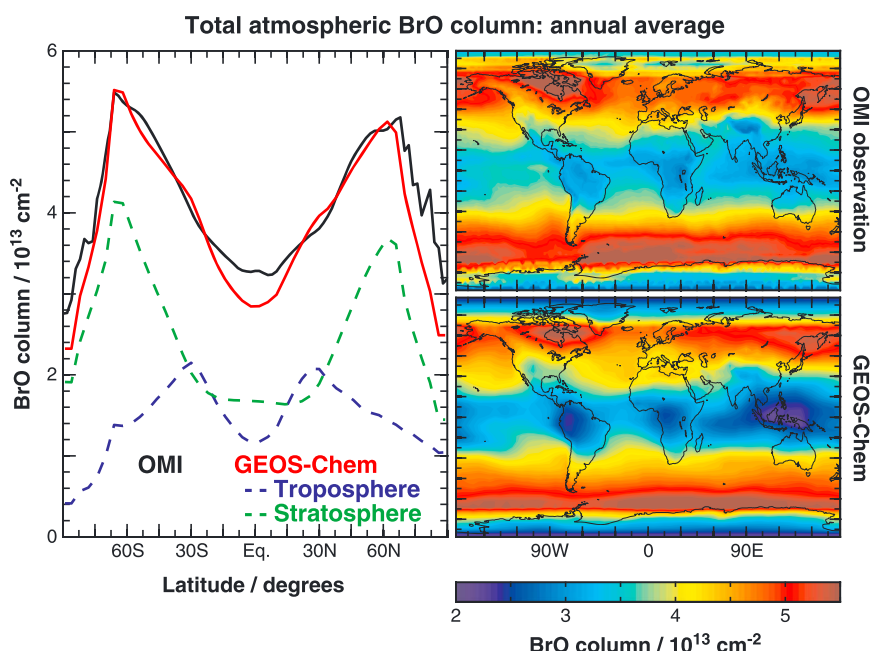


**Figure 3.** Seasonal variation of zonal mean tropospheric BrO columns in different latitudinal bands. 2007 observations from the GOME-2 satellite instrument [Theys *et al.*, 2011] are compared to GEOS-Chem values (simulation A) at the GOME-2 local overpass time (9:00–10:00). Columns obtained from the simulation including SSA dehalogenation are shown in cyan.

limit BrO concentration is smaller than 0.5 ppt, the detection limit of CU AMAX-DOAS and Ship MAX-DOAS over the open ocean during TORERO [Volkamer *et al.*, 2015], and long-path DOAS observations at Galapagos Island during CHARLEX [Gomez Martin *et al.*, 2013]. BrO increases with altitude (0.7 to 1 ppt, between 4 and 8 km) and reaches values of 1–2.5 ppt in the upper troposphere (8–13 km) [Volkamer *et al.*, 2015; Wang *et al.*, 2015].

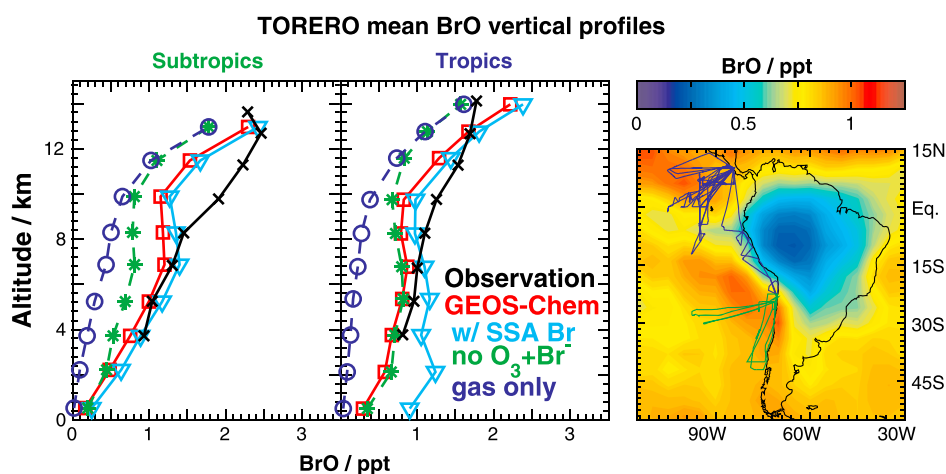
Simulation A matches closely the observed vertical profile in both the tropics and subtropics. Including SSA debromination, as in simulation B, does not change BrO in the subtropics because the model SSA in that region tends to be alkaline. However, the simulation with active sea salt dehalogenation finds about 1 ppt of BrO in tropical MBL, where SSA is simulated to be acidic. Such elevated BrO is inconsistent with the available observations [Gomez Martin *et al.*, 2013; Volkamer *et al.*, 2015]. The reason for this difference is currently not clear but points to either overestimated sources or missing bromine sinks over the Eastern Pacific ocean. High BrO in the upper troposphere includes a dominant stratospheric component as discussed in section 3 and also a 30% contribution from tropospheric multiphase chemistry involving the  $O_3 + Br^-$  reaction in ice cloud droplets and sulfate aerosol (Figure S2). The oxidation of bromide by ozone is particularly important in the upper troposphere because ozone concentrations are high and HOBr concentrations are low.

A subset of the TORERO BrO observations has recently been compared with models [Wang *et al.*, 2015]. Notably, the BrO concentration was found to be variable. Slightly lower BrO was observed during case studies with pollution influences. The BrO concentration was generally higher in the pristine free troposphere, see Figure 10 of Volkamer *et al.* [2015]. To account for atmospheric variability, Wang *et al.* [2015] compared models with average tropical and subtropical profiles in the pristine free troposphere and not individual profile



**Figure 4.** Global distribution of the total annual mean BrO column including the troposphere and stratosphere. 2007 observations from the OMI satellite instrument ([http://disc.sci.gsfc.nasa.gov/Aura/data-holdings/OMI/ombro\\_v003.shtml](http://disc.sci.gsfc.nasa.gov/Aura/data-holdings/OMI/ombro_v003.shtml)) are compared to tropospheric GEOS-Chem values (simulation A) for the OMI overpass time (13:00–14:00 local time) together with stratospheric values from GEOSCCM used as upper boundary condition for GEOS-Chem [Liang *et al.*, 2010]. The left panel separates the contributions from the troposphere and stratosphere to the total GEOS-Chem column.

case studies. Figure S7 compares the updated model results for the subset of case studies evaluated by Wang *et al.* [2015]. We find close agreement between model and observation in the subtropics. In the tropics, the model output for the selected case studies is very similar to the overall campaign average profile shown in Figure 5 but is less successful at reproducing the vertical gradient of BrO observed during these tropical case studies, as manifested by a low bias in the tropical upper troposphere (above 8 km). Wang *et al.* [2015] found



**Figure 5.** Mean vertical profiles of BrO concentration over the Southeast Pacific. Daytime aircraft observations from the TORERO campaign (January–February 2012; black cross) are compared to GEOS-Chem values (simulation A) sampled along the 17 flight tracks (red square). Also shown are results from model sensitivity simulations including SSA dehalogenation (simulation B; cyan triangle), without the aqueous phase  $O_3 + Br^-$  reaction (green asterisk), and without multiphase halogen chemistry (gas-phase chemistry only; blue circle). The observed BrO in the MBL was below the instrument detection limit of about 0.5 ppt. (right) The TORERO flight tracks superimposed on the model distribution of daytime tropospheric mean BrO mixing ratios for the flight period.

**Table 4.** Global Tropospheric Ozone Budgets in GEOS-Chem (Simulation A)<sup>a</sup>

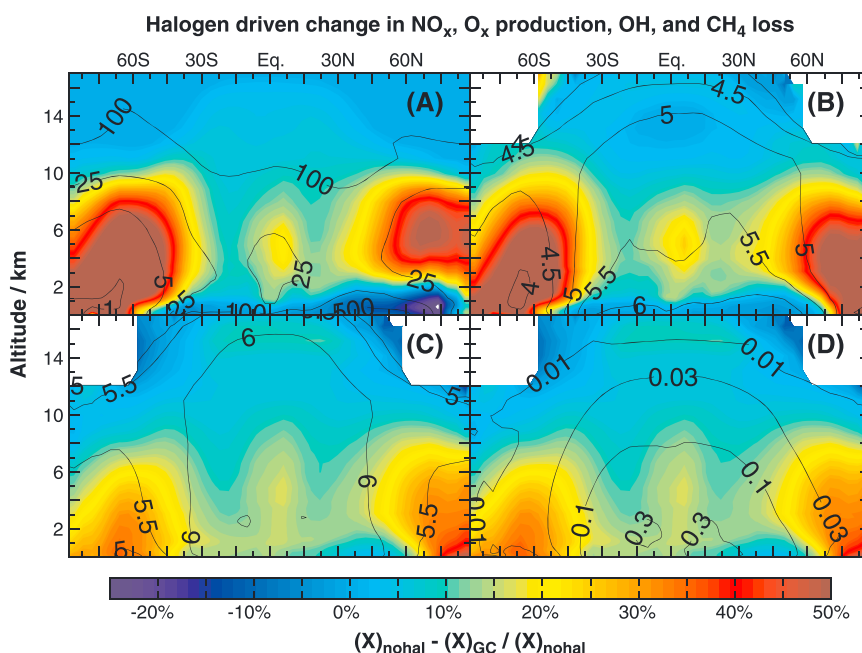
	This Study	No Halogen
Sources (Tg a <sup>-1</sup> )		
Chemistry	4964	5513
Stratosphere	515	510
Sinks (Tg a <sup>-1</sup> )		
Chemistry	4500	4926
Deposition	979	1097
Burden (Tg)	329	383
Lifetime (days)	21.9	23.2

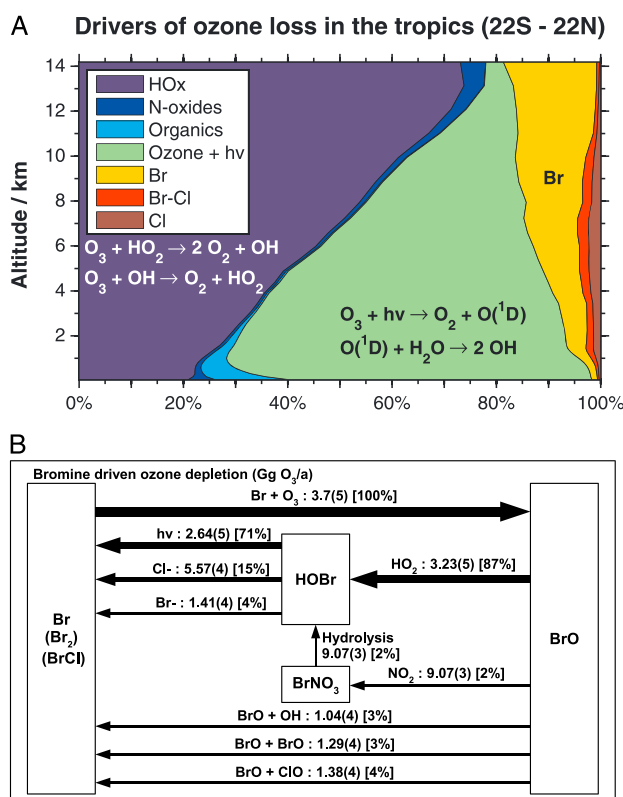
<sup>a</sup>The ozone budget is defined as that of odd oxygen (see definition of O<sub>x</sub> in text). The burden, production, and loss rates are given in O<sub>3</sub> equivalent masses. Nonozon O<sub>x</sub> contribute less than 1% of the burden.

good agreement for BrO in the lower stratosphere and noted the negligence of models to represent bromine sources from SSA in the tropical upper free troposphere. They further suggested missing multiphase chemistry involving the O<sub>3</sub> + Br<sup>-</sup> reaction during deep convective lofting of sea salt aerosols, debromination of ice cloud crystals, or overestimated wet scavenging by the model as possible reasons for the elevated BrO in the tropical upper free troposphere.

## 5. Implications for Tropospheric Ozone, OH, and Mercury

Table 4 shows the GEOS-Chem global annual budget of tropospheric ozone, with and without halogen (Br-Cl) chemistry in our simulation A, compared to previous model studies. The budget of ozone is given as that of

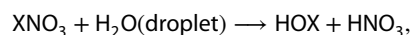
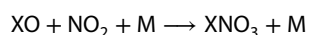




**Figure 7.** Effect of halogen chemistry on the global annual tropospheric ozone budget. The ozone budget is defined that of odd oxygen ( $\text{O}_x \equiv \text{O}_3 + \text{O} + \text{NO}_2 + 2\text{NO}_3 + \text{peroxyacynitrates} + \text{HNO}_3 + \text{HNO}_4 + 3\text{N}_2\text{O}_5 + \text{XO} + \text{HOX} + \text{XNO}_2 + 2\text{XNO}_3 + \text{Cl}_2\text{O}_2 + 2\text{OCIO}$ ) to account for rapid cycling between  $\text{O}_x$  species. (top) Relative contributions of different processes to  $\text{O}_x$  loss in the tropics (22°S to 22°N). (bottom) Pathways for bromine-catalyzed ozone loss. Rates are in  $\text{Gg Br a}^{-1}$  (Global annual average).

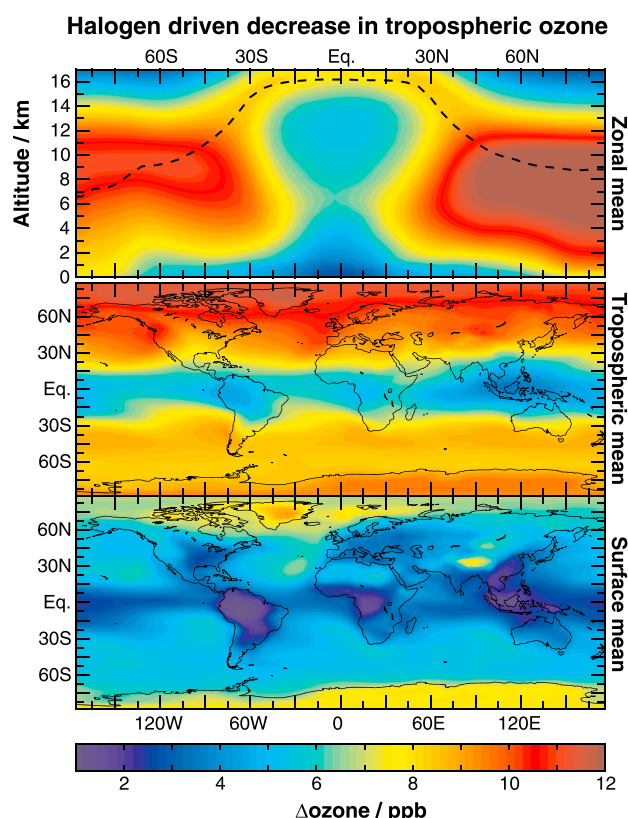
odd oxygen ( $\text{O}_x \equiv \text{O}_3 + \text{O} + \text{NO}_2 + 2\text{NO}_3 + \text{peroxyacynitrates} + \text{HNO}_3 + \text{HNO}_4 + 3\text{N}_2\text{O}_5 + \text{XO} + \text{HOX} + \text{XNO}_2 + 2\text{XNO}_3 + \text{Cl}_2\text{O}_2 + 2\text{OCIO}$ ) to account for rapid cycling between  $\text{O}_x$  species. Budget terms are close to those of Parrella *et al.* [2012] and in the range of the model literature [Wu *et al.*, 2007]. Halogen chemistry lowers the global tropospheric ozone burden by 14%, which is much larger than the 6.5% found in the previous GEOS-Chem study by Parrella *et al.* [2012]. We find that the decrease in ozone is driven by a 10% decrease in the chemical production of ozone due to lower levels of  $\text{NO}_x$  and a 5.6% decrease in the lifetime of ozone due to halogen driven catalytic ozone loss. The finding that the halogen-driven decrease in tropospheric ozone is caused by decreased  $\text{NO}_x$ -driven ozone production combined with enhanced ozone destruction is consistent with previous model studies [von Glasow *et al.*, 2004; Parrella *et al.*, 2012; Long *et al.*, 2014].

Halogen-driven  $\text{NO}_x$  loss takes place by hydrolysis of the halogen nitrates:



and decreases the global burden of tropospheric  $\text{NO}_x$  by about 6%. Hydrolysis of  $\text{BrNO}_3$  and  $\text{ClNO}_3$  contribute equally to  $\text{NO}_x$  loss. The rate of halogen-driven  $\text{NO}_x$  loss is about one fifth of the rate of  $\text{NO}_x$  loss by  $\text{NO}_2 + \text{OH}$ . The relative decrease in  $\text{NO}_x$  is largest in low- $\text{NO}_x$  areas such as the tropical free troposphere and polar regions (Figure 6). By contrast,  $\text{NO}_x$  increases in continental boundary layers. Here levels of reactive inorganic halogens are relatively low. The decrease in the level of ozone (and OH) increase the chemical lifetime of  $\text{NO}_x$  these regions.

Figure 7 summarizes the relative contributions of different processes to ozone destruction in the tropics, where most of global tropospheric ozone loss takes place [Wang *et al.*, 1998]. The relative contribution of halogen-driven ozone loss increases with altitude because of increasing  $\text{BrO}_x$  (Figure 1). Most ozone

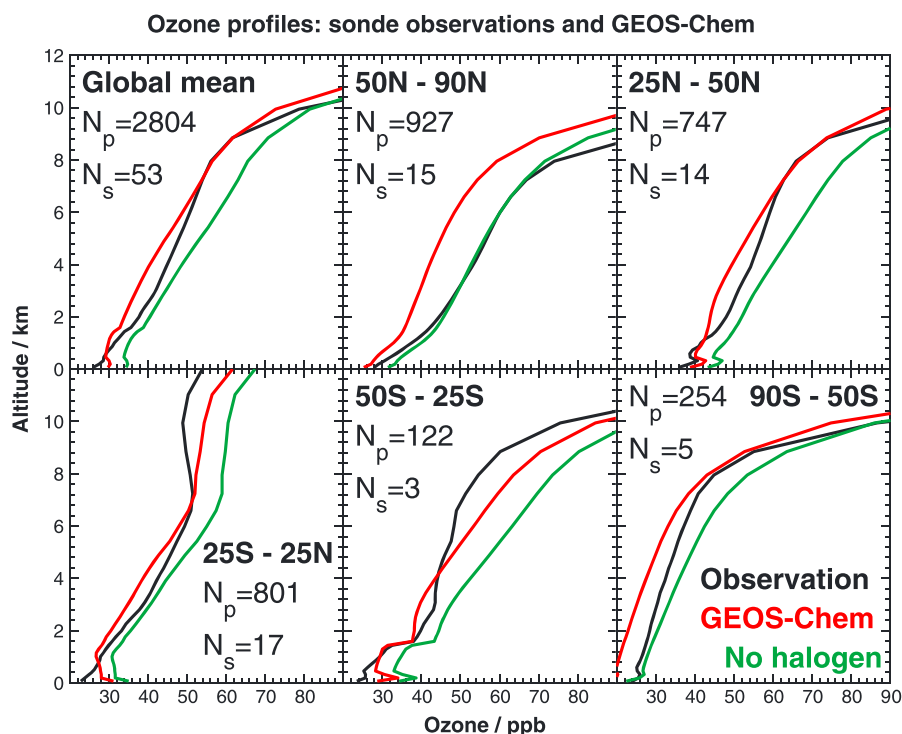


**Figure 8.** Effect of halogen chemistry on global tropospheric ozone concentrations. Values are annual mean differences between GEOS-Chem simulations not including versus including halogen chemistry:  $\Delta\text{ozone}$  is ozone from simulation with no halogen chemistry minus ozone from simulation A. (top) Zonal mean difference. (middle) Difference in tropospheric mean. (bottom) Difference in surface concentration.

destruction occurs below  $\sim 6$  km (Figure S6), where halogen chemistry drives 5–15% of ozone loss. As previously found by Parrella *et al.* [2012] and illustrated in Figure 7 (bottom), Br-catalyzed ozone loss is mainly driven by HOBr formation via the  $\text{BrO} + \text{HO}_2$  reaction (87% of total Br-catalyzed ozone loss). Chemical recycling of HOBr is important because it controls the BrO abundance. Photolysis is the dominant recycling pathway for HOBr (71%; Figure 7, bottom). The multiphase  $\text{HOBr} + \text{Cl}^-$  reaction drive about 15% of HOBr recycling, and is an example of Br-Cl synergized ozone loss.  $\text{HOBr} + \text{Br}^-$  is responsible for 4% of HOBr cycling but is critical for the cycling of bromide. The contribution of pure chlorine chemistry to chemical loss of ozone is small.

Figure 8 shows annual mean decreases in tropospheric ozone concentrations due to halogen chemistry for different regions of the troposphere. The decrease exceeds 10 ppb in the northern midlatitude free troposphere. On average, halogen chemistry decreases the tropospheric ozone column by 12% (2–7 ppb) in the tropics and 17% (6–10 ppb) at midlatitudes. The large ozone decrease in the northern high-latitude free troposphere is driven by a 20–50% decreased ozone production in this area due to lower levels of  $\text{NO}_x$  (Figure 6) and to a lesser extent decreased ozone lifetime due to halogen-driven catalytic ozone loss. Halogen chemistry lowers surface ozone between 1 and 8 ppb: The decrease is smallest in the polluted continental PBL where halogen chemistry increases ozone production due to increased levels of  $\text{NO}_x$  (Figure 6).

The predicted halogen-driven ozone decrease in the tropics falls within previous model estimates: The study of Saiz-Lopez *et al.* [2012] found halogen chemistry (including chlorine, bromine, and iodine) reduced the tropospheric ozone column by more than 10% in the tropics. Long *et al.* [2014] found halogen (chlorine and bromine) driven ozone decreases larger than 20% in the tropical MBL, and between 15 and 20% in the tropical free troposphere (FT). Wang *et al.* [2015] found that halogen (bromine and iodine) chemistry drive 34% of ozone loss over the tropical Eastern Pacific with bromine and iodine each contributing about equally. Iodine cycling has been found to contribute to halogen ozone loss in the FT and dominate in the MBL [Saiz-Lopez *et al.*, 2012; Dix *et al.*, 2013; Wang *et al.*, 2015; Sherwen *et al.*, 2016]. The GEOS-Chem-based study of

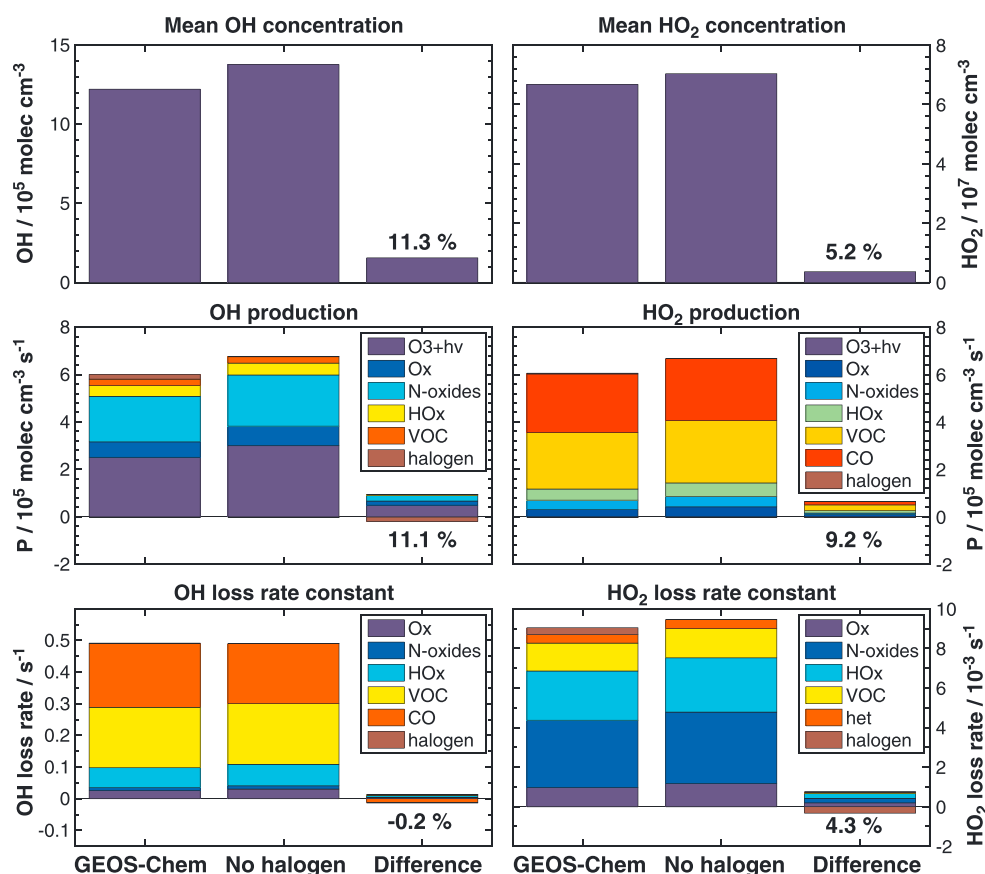


**Figure 9.** Annual average vertical profile from ozonesonde observations for five zonal bands compared to model ozone from the current simulation and a simulation with no halogen chemistry.  $N_s$  and  $N_p$  are the number of sites and profiles, respectively, included in the averaging. The location of the sites is given in the supporting information.

Sherwen *et al.* [2016] found a 9% decrease in the global burden of ozone by introducing iodine chemistry and a halogen (bromine and iodine) driven loss comparable to this work of 14.4%. Iodine drove a  $750 \text{ Tg a}^{-1}$  increase in the  $\text{O}_x$  loss rate and had the greatest proportional effect in the MBL and upper troposphere. Sherwen *et al.* [2016] used the bromine simulation detailed in Parrella *et al.* [2012] which results in smaller decrease in tropospheric ozone burden (6.5%) than the 14% from chlorine and bromine presented here. The increased BrO concentrations calculated in this work may impact the Sherwen *et al.* [2016] simulation through the coupling reaction  $\text{BrO} + \text{IO}$ . However, Sherwen *et al.* [2016] notes that the coupling in their simulation is weak. Overall, it appears that both bromine and iodine have significant impacts on tropospheric composition and a coupled Cl-Br-I simulations with state of the science representations of the emissions, processing, and deposition of the halogens is required to evaluate the full impact.

Previous studies have shown clear evidence for bromine-catalyzed ozone loss in observations of surface ozone in Arctic spring [Barrie *et al.*, 1988] and over the ocean [Read *et al.*, 2008]. Evidence of halogen chemistry in free tropospheric ozone observations is more elusive because the chemical time scales are longer and effects may be masked by errors in transport, lightning  $\text{NO}_x$  emissions, or other chemical aspects. Figure 9 compares GEOS-Chem (simulation A) with and without halogen chemistry to mean ozonesonde observations in five latitudinal band (L. Hu *et al.*, manuscript in preparation, 2016). Halogen chemistry lowers ozone and improves agreement with observation in the tropics and at midlatitudes. At Northern high latitudes ( $>50^\circ\text{N}$ ) the no halogen chemistry-simulated ozone profile shows very close agreement with observations and the simulation with halogen chemistry is biasedly low.

We find that halogen chemistry lowers global tropospheric  $\text{HO}_x$  concentrations by 5% and OH by 11% (Figure 10). The effect on OH is much larger than the 4% decrease predicted by Parrella *et al.* [2012]. The decrease in  $\text{HO}_x$  largely follows from the decrease in ozone. In addition, halogen chemistry decreases the OH/ $\text{HO}_2$  ratio by  $\sim 10\%$  in the tropics and  $\sim 20\%$  at high latitudes: Decreased levels of  $\text{NO}_x$  results in a diminished OH source from  $\text{HO}_2 + \text{NO}$ , and the added OH source from HOBr photolysis is too small to compensate. The model study of Long *et al.* [2014] also found halogen chemistry to decrease the OH/ $\text{HO}_2$  ratio. The relative decrease in OH is largest at high latitudes ( $\sim 30\%$ ; Figure 6c); however, the absolute change in OH between the



**Figure 10.** The global annual average GEOS-Chem budget of HO<sub>x</sub> (OH and HO<sub>2</sub>) simulated with (simulation A) and without halogen chemistry.

simulation with and without halogen peaks in the tropical free troposphere. The lower OH concentration in our simulation prolongs the lifetimes of methane against tropospheric oxidation by OH by 16% to 9.5 years, better matching the current best observational estimate of  $11.2 \pm 1.3$  years [Prather et al., 2012].

Bromine chemistry may also have implications for the atmospheric lifetime and deposition of mercury. The Br atom, which cycles photochemically with BrO, is thought to be the main atmospheric oxidant of Hg(0) [Goodsite et al., 2004; Holmes et al., 2006, 2010]. Using the Hg oxidation mechanism of Holmes et al. [2006] and the Br fields of our simulation A, we estimate a global mean atmospheric lifetime of 6 months for Hg(0) against oxidation to Hg(II) by Br. This is consistent with observations of Hg(0) atmospheric gradients implying a lifetime of 0.5–2 years against deposition [Slemr et al., 1985], which would be longer than the lifetime of Hg(0) against oxidation if atmospheric Hg(II) reduction takes place [Holmes et al., 2010]. A major difference between our results and those of Holmes et al. [2010] and Parrella et al. [2012] is that our simulated Br concentrations are much lower over the Southern Ocean due to suppression of Br<sub>y</sub> release from sea salt, so that mercury deposition in that region is lower. Using the more recent Dibble et al. [2012] mechanism as implemented in GEOS-Chem by H. M. Horowitz et al. (manuscript in preparation, 2016) with our halogen fields (Br, BrO, Cl, and ClO), we find a global atmospheric lifetime of 3 months for Hg(0) against oxidation to Hg(II).

## 6. Conclusion

We used recent aircraft observations of BrO in the tropical and subtropical atmosphere, together with satellite observations, to improve our understanding of global tropospheric bromine chemistry and its implications. The observations point to higher background concentrations of BrO in the free troposphere than have been simulated by past models.

Sustaining the high levels of BrO seen in the observations requires rapid recycling of bromine radicals ( $\text{BrO}_x \equiv \text{Br} + \text{BrO}$ ) from the pool of nonradical  $\text{Br}_y$  reservoirs. This can be achieved through multiphase chemistry in aerosols and clouds. We followed recommendations from IUPAC [Ammann *et al.*, 2013] by implementing into the GEOS-Chem global chemical transport model (CTM) an ensemble of multiphase halogen reactions including oxidation of  $\text{HBr}/\text{Br}^-$  by  $\text{HOBr}$ ,  $\text{ClNO}_3$ , and ozone in aerosols as well as liquid and ice clouds. This involved the addition of tropospheric chlorine chemistry in GEOS-Chem, enabling a broader examination of coupled Br-Cl chemistry. It also involved coupling of bromine chemistry to a dynamic simulation of sea salt aerosol (SSA) alkalinity since release of SSA bromide requires an acidified aerosol. The SSA-driven source of bromine radicals was not included in the standard simulation (A) but was examined in a sensitivity simulation (B).

Results from the updated GEOS-Chem simulation indicate a global mean tropospheric BrO concentration of about 1 ppt in the daytime. Most of this BrO is in the free troposphere where its main source is from oxidation and photolysis of marine biogenic bromoform ( $\text{CHBr}_3$ ). Stratospheric input also contributes in the upper troposphere. The dominant bromine reservoirs are  $\text{HOBr}$  in the daytime and  $\text{BrCl}$  at night, reflecting the rapid recycling of  $\text{HBr}$  in cloud and sulfate aerosol. Debromination of SSA improves BrO observations in the marine boundary layer over the Western Pacific but presents a challenge to reconcile model predictions with observations of low BrO in the marine boundary layer over the tropical Eastern Pacific.

Our simulated BrO concentrations are consistent with GOME-2 and OMI satellite column observations in the tropics and midlatitudes and corroborate cloud-slicing analysis and aircraft observations, indicating that most tropospheric BrO is in the free troposphere. We find that the troposphere accounts for over half of the total (troposphere + stratosphere) BrO column in the subtropics. The model reproduces the increase of BrO with altitude in TORERO aircraft observations over the subsiding Southeast Pacific.

Halogen (bromine-chlorine) chemistry, as implemented in our simulation, decreases the global burden of tropospheric ozone in GEOS-Chem by 14% (2–7 ppb in the tropics and 6–10 ppb at midlatitudes). This reflects a decrease in chemical production of ozone by 10% (due to halogen-driven  $\text{NO}_x$  loss) and a shortening of the ozone lifetime by 5.6% (due to Br-catalyzed ozone loss). In most of the troposphere, halogen chemistry decreases  $\text{NO}_x$  levels through formation of halogen nitrates followed by hydrolysis to  $\text{HNO}_3$ . In polluted boundary layers, however, halogen chemistry increases  $\text{NO}_x$  levels by decreasing OH. Thus, halogen chemistry decreases mean surface ozone concentrations in the North America by only 2–6 ppb, and increases the relative importance of regional pollution from domestic  $\text{NO}_x$  sources versus transported background ozone. Halogen chemistry lowers global mean tropospheric OH concentrations by 11%, driven by decreases in both ozone and  $\text{NO}_x$ . The resulting methane lifetime against oxidation by tropospheric OH in GEOS-Chem increases by 16% to 9.5 years, improving the agreement with current best estimates. The global mean atmospheric lifetime of elemental mercury  $\text{Hg}(0)$  against oxidation to  $\text{Hg}(\text{II})$  by Br atoms is 6 months, within the observational constraints. Alkaline SSA over the Southern Ocean implies that Br atom concentrations should be relatively low there, suppressing  $\text{Hg}(\text{II})$  deposition.

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