Methyl iodide: Atmospheric budget and use as a tracer of marine convection in global models

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We simulate the oceanic and atmospheric distribution of methyl iodide (CH$_3$I) with a global 3-D model driven by assimilated meteorological observations from the Goddard Earth Observing System of the NASA Data Assimilation Office and coupled to an oceanic mixed layer model. A global compilation of atmospheric and oceanic observations is used to constrain and evaluate the simulation. Seawater CH$_3$I(aq) in the model is produced photochemically from dissolved organic carbon, and is removed by reaction with Cl$^-$ and emission to the atmosphere. The net oceanic emission to the atmosphere is 214 Gg yr$^{-1}$. Small terrestrial emissions from rice paddies, wetlands, and biomass burning are also included in the model. The model captures 40% of the variance in the observed seawater CH$_3$I(aq) concentrations. Simulated concentrations at midlatitudes in summer are too high, perhaps because of a missing biological sink of CH$_3$I(aq). We define a marine convection index (MCI) as the ratio of upper tropospheric (8–12 km) to lower tropospheric (0–2.5 km) CH$_3$I concentrations averaged over coherent oceanic regions. The MCI in the observations ranges from 0.11 over strongly subsiding regions (southeastern subtropical Pacific) to 0.40 over strongly upwelling regions (western equatorial Pacific). The model reproduces the observed MCI with no significant global bias (offset of only +11%) but accounts for only 15% of its spatial and seasonal variance. The MCI can be used to test marine convection in global models, complementing the use of radon-222 as a test of continental convection.

INDEX TERMS: 0312 Atmospheric Composition and Structure: Air/sea constituent fluxes (3339, 4504); 0322 Atmospheric Composition and Structure: Constituent sources and sinks; 0368 Atmospheric Composition and Structure: Troposphere—constituent transport and chemistry; KEYWORDS: methyl iodide, marine convection, atmospheric tracer, global budget of methyl iodide


1. Introduction

Methyl iodide (CH$_3$I) is emitted to the atmosphere by the oceans and photolyzes with a lifetime of the order of a week. Aircraft measurements over the past decade have provided large data sets of CH$_3$I concentrations over the oceans up to 12 km altitude [Blake et al., 1999, and references therein]. These observations have been used to constrain vertical mixing rates in 1-D models of atmospheric transport [Davis et al., 1996; Wang et al., 2000, 2002] and to diagnose outflow from marine convection in the upper troposphere [Cohan et al., 1999; Staudt et al., 2002]. Beyond its value as a tracer of transport, CH$_3$I is also of interest in the upper troposphere and lower stratosphere as a source of iodine radicals for ozone destruction [Solomon et al., 1994; Davis et al., 1996]. In addition, methyl iodide is a leading candidate to replace methyl bromide as a soil fumigant [Waggoner et al., 2000].

We present here a global 3-D simulation of CH$_3$I in the ocean-atmosphere coupled system and use a large ensemble of worldwide atmospheric and oceanic CH$_3$I observations to evaluate model results. Our first objective is to develop CH$_3$I as a tracer for testing marine convection in global 3-D atmospheric models, complementing the standard use of $^{222}$Rn as a tracer for continental convection [Jacob et al., 1997]. Our second objective is to quantify better the global atmospheric budget of CH$_3$I.
Table 1. Global CH₃I Emission Estimates in the Literature

<table>
<thead>
<tr>
<th>Source</th>
<th>Global Emission, Gg/yr</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oceanic*</td>
<td>270</td>
<td>Liu and Slater [1974]</td>
</tr>
<tr>
<td></td>
<td>1300</td>
<td>Rasmussen et al. [1982]</td>
</tr>
<tr>
<td>300–500</td>
<td>Singh et al. [1983]</td>
<td></td>
</tr>
<tr>
<td>140</td>
<td>Nightingale [1991]</td>
<td></td>
</tr>
<tr>
<td>800</td>
<td>Reifenhauser and Heumann [1992]</td>
<td></td>
</tr>
<tr>
<td>150</td>
<td>Campos et al. [1996]</td>
<td></td>
</tr>
<tr>
<td>130–350</td>
<td>Moore and Groszko [1999]</td>
<td></td>
</tr>
<tr>
<td>Terrestrial</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Biomass burning</td>
<td>&lt;10</td>
<td>Andreea et al. [1996]</td>
</tr>
<tr>
<td>Plant-soil systems</td>
<td>3.4</td>
<td>Blake et al. [1996b]</td>
</tr>
<tr>
<td>Rice paddies</td>
<td>20</td>
<td>Muramatsu and Yoshida [1995]</td>
</tr>
<tr>
<td></td>
<td>71</td>
<td>Redeker et al. [2000]</td>
</tr>
<tr>
<td>Peatland ecosystems</td>
<td>1.4</td>
<td>Dimmer et al. [2000]</td>
</tr>
<tr>
<td>Wetlands</td>
<td>7.3</td>
<td>Dimmer et al. [2000]</td>
</tr>
</tbody>
</table>

*Extrapolated from field measurements over limited regions.

[4] A simple way to implement CH₃I as a tracer of vertical transport in atmospheric models is to use observed marine boundary layer (MBL) concentrations as a lower boundary condition and simulate the vertical gradient in the tropospheric column. The 1-D models cited above used this approach to constrain their mean vertical transport rates. The approach is less attractive in a time-dependent 3-D approach to constrain their mean vertical transport rates. A boundary condition at the oceanic emission estimates in Table 1 [1984].

[5] Data for oceanic concentrations of CH₃I are available from a number of ship cruises. Extrapolations from individual regional data sets have led to global emission estimates in the literature ranging from 130 to 1300 Gg yr⁻¹ (Table 1). Consideration of the global data base of CH₃I(aq) concentration measurements allows a more constrained estimate, as described in this paper. Biological sources from algae and phytoplankton have been examined in a number of laboratory studies and could contribute to elevated CH₃I in coastal environments, but they appear to be negligible on a global scale when compared to the oceanic emission estimates in Table 1 [Manley and Dastoor, 1987, 1988; Nightingale et al., 1995; Manley and de la Cuesta, 1997; Giese et al., 1999]. A laboratory study by Moore and Zafiriou [1994] indicates that photochemical degradation of dissolved organic carbon (DOC) could possibly provide the main oceanic source of CH₃I(aq). Field measurements by Happell and Wallace [1996] also suggest that the source of CH₃I(aq) is photochemical. The simulation of CH₃I(aq) in our global model is more consistent with a photochemical source than with a biological source, as described below.

[6] The main sinks of CH₃I(aq) in the oceans are thought to be emission to the atmosphere and nucleophilic Sₘ₂ reaction with Cl⁻. These two sinks are of comparable magnitude [Zafiriou, 1975]; either may dominate depending on surface wind speed (which affects emission to the atmosphere) and seawater temperature (which affects the rate of reaction with Cl⁻). Hydrolysis is typically an order of magnitude slower than reaction with Cl⁻ [Moelwyn-Hughes, 1938]. Photoysis is unimportant [Zika et al., 1984].

[7] Little is known regarding continental sources of CH₃I (Table 1). There is a source from rice paddies [Muramatsu and Yoshida, 1995] which Redeker et al. [2000] have estimated to be as large as 71 Gg yr⁻¹ globally. There is a small CH₃I source from biomass burning of 3–10 Gg yr⁻¹ [Andreae et al., 1996; Blake et al., 1996b]. Dimmer et al. [2000] have presented evidence for a small source from peatland ecosystems. The continental source of CH₃I may be much less than the oceanic source on a global scale, but measurements of atmospheric concentrations of CH₃I over the continents are too few at present to be illuminating.

2. Model Description

2.1. General

[8] We use the GEOS-CHEM global 3-D model of tropospheric chemistry [Bey et al., 2001a; Liu et al., 2001] driven by assimilated meteorological observations from the Goddard Earth Observing System (GEOS) of the NASA Data Assimilation Office (DAO). Allen et al. [1996a] and Liu et al. [2001] previously used simulations of ²²⁴Rn with the GEOS fields to test the representation of continental convection. The GEOS-CHEM model has been applied to a range of tropospheric chemistry problems [Bey et al., 2001a, 2001b; Fiore et al., 2002; Jacob et al., 2002; Li et al., 2000, 2001, 2002; Liu et al., 2001, 2002; Martin et al., 2002a, 2002b; Palmer et al., 2001].

[9] In the present application we use GEOS fields for 1994 available with 2° latitude × 2.5° longitude horizontal resolution and 20 sigma levels in the vertical extending up to 10 hPa. We degrade the horizontal resolution to 4° × 5° for computational expediency. The lowest 3 km of the atmosphere are resolved by 6 levels centered at 0.15, 0.35, 0.71, 1.2, 1.9, and 2.6 km above the surface for an air column based at sea level. Advection is computed every 30 min with a flux-form semi-Lagrangian method [Lin and Rood, 1996]. Moist convection is computed using the GEOS convective, entrainment, and detrainment mass fluxes as described by Allen et al. [1996a, 1996b]. We assume full mixing within the GEOS-diagnosed atmospheric mixed layer generated by surface instability. The simulation is conducted for 18 months (July 1993 through December 1994) starting from low CH₃I concentrations as initial conditions. The first 6 months are used to achieve initialization and we focus our attention on the 12-month simulation for 1994. Methyl iodide emitted from different sources (ocean, biomass burning, wood fuel, rice paddies, and wetlands), is carried as different tracers in the model.

[10] The atmospheric CH₃I sink from photolysis is calculated in GEOS-CHEM using the FAST-J algorithm of Wild et al. [2000], implemented as described by Bey et al. [2001a]. Surface albedos and vertically resolved cloud optical depths are taken from the GEOS data with 6-hour resolution. CH₃I absorbs over the wavelength range 300–340 nm and we use absorption cross sections from Roehl et al. [1997]. We assume unit quantum yield as has been observed for CH₂ClI and CH₂I₂ in that wavelength range [Roehl et al., 1997, and references therein]. The resulting lifetime of CH₃I is about 4 days in the tropical troposphere (Figure 1). Oxidation of CH₃I by OH is a negligibly small
sink, amounting to only about 1% of photolysis [Brown et al., 1990].

2.2. Oceanic Emission of Methyl Iodide

[11] Oceanic emission of CH₃I to the atmosphere is computed following Liss and Slater [1974],

\[ \Phi = k ([CH₃I(aq)] - KH[CH₃I(g)]). \]  

(1)

Here \( \Phi \) is the net flux of CH₃I from the ocean to the atmosphere representing a balance between emission from the ocean (\( \Phi_{em} = k(CH₃I(aq)) \)) and deposition from the atmosphere (\( \Phi_{dep} = -kK_H(CH₃I(g)) \)) and \( k \) is the sea-to-air transfer velocity. \( K_H \) is the dimensionless Henry’s law constant for CH₃I defined as the ratio of aqueous to gas-phase concentrations at equilibrium (\( K_H = 3.4 \) at 298 K and 1 atm, \( \Delta H_{298}/R = -4300 \) K available at http://www.mpe-mainz.mpg.de/sander/res/res/henry.html). We parameterize \( k \) as a function of the surface wind speed following Nightingale et al. [2000], who found from tracer experiments an optimal wind speed dependence intermediate between the previous results of Liss and Merlivat [1986] and Wanninkhof [1992].

[12] Using a box model for the oceanic mixed layer of depth \( z \) and assuming no horizontal flux divergence, we write a steady state expression for the local concentration of CH₃I(aq),

\[ [CH₃I(aq)] = \frac{P + k/zK_H[CH₃I(g)]}{(k/z) + k_C[Cl^-]}, \]  

(2)

where \( P \) is the oceanic production rate per unit volume (ng L⁻¹ h⁻¹), and \( k_C \) is the rate constant for reaction with Cl⁻. We use \( k_C = 7.78 \times 10^{13} \exp[-13518/\mathcal{T}] \) M⁻¹ s⁻¹ from Elliott and Rowland [1993], who provide the only temperature-dependent data for this reaction. The temperature dependence is very strong; an increase in ocean mixed layer temperature (MLT) of 20 K as observed between polar latitudes and the tropics corresponds to an approximately twenty-fold increase in \( k_C \).

In our simulation we use monthly mean MLT fields from the data set of Woodruff et al. [1987] and a uniform [Cl⁻] = 0.54 M. For typical values \( k = 10 \) cm h⁻¹, \( z = 50 \) m, and \( k_C = 1.2 \times 10^{-13} \) h⁻¹ (where MLT = 292 K), the lifetimes of CH₃I(aq) against transfer to the atmosphere and reaction with Cl⁻ are 21 and 35 days, respectively, for an overall lifetime of 13 days. Hence ventilation and chemical reaction with Cl⁻ make comparable contributions to the CH₃I(aq) sink depending on local wind speed and MLT [Zafiriou, 1975].

[14] For our model, we specify the atmospheric concentration \([CH₃I(g)]\) in equations (1) and (2) with the local value from the previous time step of the GEOS-CHEM simulation, so that the oceanic and atmospheric simulations are fully coupled. The CH₃I(aq) production rate \( P \) is the effective lower boundary condition of the model. Following the laboratory experiments of Moore and Zafiriou [1994] and the supporting observational evidence of Happell and Wallace [1996], we assume that the production of CH₃I(aq) in the open oceans is driven by a photochemical mechanism dependent on the supply of dissolved organic carbon (DOC). As a crude parameterization, we scale the production rate as the product of the solar radiation flux at the surface (RAD) and the dissolved organic carbon concentration (DOC),

\[ P = \alpha (RAD \times DOC), \]

(3)

where \( \alpha \) is a scaling parameter. Monthly average values of RAD are taken from the GEOS fields for 1994, and monthly average fields of DOC are taken from the oceanic GCM of Six and Maier-Reimer [1996]. A value \( \alpha = 0.1 \) m² W⁻¹ h⁻¹ was obtained by least-squares fit of model results to observed seawater CH₃I(aq) concentrations, as described in section 3. The resulting \( r^2 \) is 0.4. A sensitivity simulation where \( P \) was scaled to net primary productivity (NPP) to simulate a biological source of CH₃I(aq) was found to have no success in reproducing the observed distributions of CH₃I(aq).

[15] The oceanic emission of CH₃I as computed by the model is thus determined by MLT, RAD, DOC, and wind speed (Figure 2). The model predicts maximum emission at midlatitudes in the spring-summer hemisphere where solar radiation is high and MLT is relatively low. DOC is less variable than RAD in space and time but shows subtropical maxima that contribute to the high CH₃I fluxes there. The annual net emission flux to the atmosphere in the model is 214 Gg yr⁻¹ (Table 2), comparable to the median (270 Gg yr⁻¹) of previous estimates (Table 1) but at the low end of the range (130–1300 Gg yr⁻¹).

2.3. Terrestrial Emission of Methyl Iodide

[16] We include emission of CH₃I from biomass burning and wood fuel by applying a CH₃I/CO emission ratio of 4.0 × 10⁻⁶ vol/vol [Ferek et al., 1998] to a CO emission inventory with 1° × 1° spatial resolution and monthly temporal resolution (J.A. Logan, personal communication, 2000). The recent review of Andréae and Merlet [2001] gives CH₃I/CO emission ratios of 1.5 × 10⁻⁶ vol/vol, 1.3 × 10⁻⁶ vol/vol, and 1.1 × 10⁻⁶ vol/vol for savanna, tropical forest, and extratropical forest, respectively.
Global CH$_3$I emissions from rice paddies and natural wetlands are taken to be 71 Gg yr$^{-1}$ and 7.3 Gg yr$^{-1}$, respectively, following the estimates of Redeker et al. [2000] and Dimmer et al. [2000]. These emissions are distributed spatially following the corresponding patterns of CH$_4$ emission from the inventory of Fung et al. [1991]. Peatland ecosystems have been estimated to provide an additional 1.4 Gg yr$^{-1}$ source of CH$_3$I [Dimmer et al., 2000] but this is neglected in the model. Muramatsu and Yoshida [1995] estimated a global source from rice paddies of 20 Gg yr$^{-1}$, much less than the estimate of Redeker et al. [2000]. As we will see in section 4.1, it appears that the Redeker et al. [2000] estimate may be too high.

Table 3 shows the annual emission flux of CH$_3$I to the atmosphere in the model, where each of the five sources is represented by a separate tracer. The ocean accounts for 70% of the global source. Emissions from rice paddies provide a sizeable fraction (24%) of the global budget of CH$_3$I and are highly localized in southeast Asia.

### Table 3. Global Atmospheric Budget of CH$_3$I in the Model

<table>
<thead>
<tr>
<th>Sources</th>
<th>Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ocean$^a$</td>
<td>214 Gg yr$^{-1}$</td>
</tr>
<tr>
<td>Rice paddies</td>
<td>71 Gg yr$^{-1}$</td>
</tr>
<tr>
<td>Wetlands</td>
<td>7 Gg yr$^{-1}$</td>
</tr>
<tr>
<td>Biomass burning</td>
<td>9 Gg yr$^{-1}$</td>
</tr>
<tr>
<td>Wood fuel</td>
<td>3 Gg yr$^{-1}$</td>
</tr>
<tr>
<td>Total source</td>
<td>304 Gg yr$^{-1}$</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Sink, Gg yr$^{-1}$</th>
<th>304</th>
</tr>
</thead>
<tbody>
<tr>
<td>Photolysis</td>
<td>304</td>
</tr>
</tbody>
</table>

### Additional Parameters

<table>
<thead>
<tr>
<th>Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Inventory, Gg</td>
</tr>
<tr>
<td>Lifetime, days</td>
</tr>
</tbody>
</table>

$^a$ Net source representing a balance between emission to the atmosphere and atmospheric deposition to the ocean (see Table 2).


Peninsula [Reifenhauser and Heumann, 1992] and the Peru coast [Singh et al., 1983].

[20] Simulated (monthly mean) and observed CH$_3$I(aq) concentrations are correlated with $r^2 = 0.4$ (Figure 4). The model predicts maximum oceanic concentrations at subtropical to middle latitudes in the summer season where solar radiation is high, but MLT is relatively low. Similar subtropical maxima are seen in some of the cruises [Schall
et al., 1997; Moore and Groszko, 1999]. Most of the fine structure in the model distribution is driven by the variability in MLT. Observations are higher in the eastern than in the central Pacific, a feature reproduced by the model and the possibility in MLT. Observations are higher in the eastern than in the central Pacific, a feature reproduced by the model and the results indicate a similar seasonal amplitude and phase.

The model is more successful in the Pacific than the Atlantic. It generally overestimates concentrations in the Atlantic by about a factor of 2. Most severe is the model overestimate in the Labrador Sea in spring and summer, where simulated concentrations are 1–2 orders of magnitude higher than observations from two separate cruises [Moore and Tokarczyk, 1993; Moore and Groszko, 1999]. Moore and Groszko [1999] noted no correlation with solar radiation in their Labrador Sea data.

More generally, it appears that the model systematically overestimates observations at middle to high latitudes in spring-summer where solar radiation is high and MLT is low. Under these conditions the CH$_3$I(aq) loss rate in the model is determined by ventilation to the atmosphere. In order to match observations, one would need an additional loss process with a rate constant of the order of $1 \times 10^{-2}$ h$^{-1}$. Biological degradation of CH$_3$I in seawater or high-latitude upwelling could possibly provide this missing sink but no data are available. A parallel can be drawn to methyl bromide (CH$_3$Br), for which measurements in the open ocean surface waters of the North Atlantic indicate a rate constant for biological degradation that can reach $1.2 \times 10^{-2}$ h$^{-1}$ [Tokarczyk and Saltzman, 2001]. In high-latitude ocean waters a loss rate of unknown origin of $2.4 \times 10^{-3}$ h$^{-1}$ was required to explain the observed CH$_3$Br subsaturation [Moore and Webb, 1996; Lobert et al., 1997].

The NOAA/CMDL BLAST cruises in the eastern Pacific (BLAST 1), the Atlantic (BLAST 2), and the Antarctic (BLAST 3), provide CH$_3$I mixing ratio data for both surface air and seawater [Lobert et al., 1995, 1996; J. Butler, unpublished data, 1999]. These data allow determination of the saturation ratio $S = \Phi_{\text{aer}}/\Phi_{\text{sea}} = [\text{CH}_3\text{I(aq)}]/(K_d[\text{CH}_3\text{I(g)}])$. The resulting values of $S$ are shown in Figure 5 and compared to the GEOS-CHEM model results. Since the calibration problems associated with the Blast 1 CH$_3$I data affect both the air and seawater values, the offset cancels out in the calculation of $S$, and these values are included in Figure 5. Observed saturation ratios are of the order of 50–100 at 30$^\circ$S–30$^\circ$N and decrease at higher latitudes. The model captures the magnitude and latitudinal variability in the saturation ratio (Figure 5). In the model, the decrease in $S$ at higher latitudes is driven by larger atmospheric CH$_3$I concentrations due to a longer CH$_3$I lifetime.

Although $S$ is low for the BLAST 3 cruise carried out at southern latitudes in austral fall, it is greater than unity, indicating that the Southern Ocean is a net source of CH$_3$I. Previously, Happell and Wallace [1996] measured subsaturation of CH$_3$I in the Greenland and Norwegian Seas in November. The model does not find subsaturation in this region, despite close agreement between the model and observed oceanic concentrations of CH$_3$I (0.05 ng/L). The observed atmospheric mixing ratio (2.4 pptv) is much higher than the corresponding model value (0.6 pptv).

We conducted sensitivity simulations to investigate the possible role of biological production as a source of CH$_3$I(aq) to the oceans but results were negative. Scaling the CH$_3$I(aq) production rate $P$ to net primary productivity (NPP) fields (taken from Six and Maier-Reimer [1996]) rather than to DOC $\times$ RAD in the model resulted in no significant correlation between simulated CH$_3$I(aq) fields and the observations of Figure 3. Addition of a biological source scaled to NPP to the photochemical source used in the standard simulation led to a systematic degradation of the comparison between simulated and observed CH$_3$I(aq) concentration fields.

4. Atmospheric Methyl Iodide Distribution

Our compilation of atmospheric CH$_3$I observations comprises six ship cruises and eight aircraft missions (Table

![Figure 5](https://example.com/figure5.png)  
*Figure 5.* CH$_3$I seawater saturation ratio $S = [\text{CH}_3\text{I(aq)}]/(K_d[\text{CH}_3\text{I(g)}])$ calculated from the BLAST cruises (Figure 3), as a function of latitude. The corresponding GEOS-CHEM results are shown for comparison.
Data from the eight aircraft missions were divided into 21 regions (Figure 6) chosen for their relatively homogeneous air mass characteristics and for conformity with the regions previously selected by Emmons et al. [2000] for averaging of aircraft atmospheric chemistry observations. The aircraft data are mostly over the Pacific and are clustered in the September–October and February–March periods. Data from Blast 1 (Pacific Ocean, January–February 1994) are not included because a calibration offset results in an overestimation of the atmospheric mixing ratios.

### 4.1. Marine Boundary Layer

[27] Figure 6 shows the geographical distribution of boundary layer (0–1 km) concentrations for the two seasons. Comparison of ship with collocated aircraft (0.2–1 km) measurements show that the latter are on average 50% lower. Although the discrepancy might be due to a calibration difference, it could also reflect a vertical gradient of CH$_3$I concentrations between the surface layer (sampled by the ships) and the MBL (sampled by aircraft). Such a gradient between the surface layer and the MBL has been previously noted in the case of DMS between concurrent ship and aircraft measurements during the ACE-1 mission [Suhre et al., 1998; Mari et al., 1998]. The discrepancy is inconsequential for the model calculation of the sea-to-air flux of CH$_3$I since $S > 1$ under almost all conditions.

[28] The right-hand panels of Figure 6 show the distributions of atmospheric CH$_3$I in the GEOS-CHEM model. For comparison with the aircraft regions, the model data have been averaged between 0 and 1 km, and for comparison with the ship observations the model data are for the lowest model layer (0–0.15 km). The model overestimates observed mixing ratios by a factor of 3 over the Southern Ocean and Tasman Sea (regions 19 and 20) during the austral summer, suggesting an overestimated source or a missing seasonal sink, perhaps biological, for CH$_3$I in the ocean; such a sink was proposed previously in section 3 from comparison of simulated and observed CH$_3$I(aq) concentrations. The high concentrations off the coast of China in the model summer reproduce qualitatively the same feature in the observations but are too high by a factor of 2. Rice paddies are the dominant source of CH$_3$I for this region in the model, and the overestimate suggests that the rice paddy source of CH$_3$I from Redeker et al. [2000] is too high.

[29] It is evident from Figure 6 that the model has difficulty reproducing the patterns observed in the CH$_3$I MBL concentration field. This problem does not necessarily compromise the use of the relative CH$_3$I vertical profiles as a test of convection, as demonstrated in the following section.

### 4.2. Vertical Profiles

[30] Observed vertical profiles of CH$_3$I concentrations for selected regions of Figure 6 are compared to model results in Figure 7. In most cases the simulated concentrations are dominated by the oceanic source. When another source contributes more than 5% to the total sum, it is explicitly indicated. The rice paddy source makes a 30% contribution to the CH$_3$I concentration in the upper troposphere over the North Pacific and Hawaii during PEM-West A and ACE 1. Biomass burning accounts for 50% of the boundary layer concentration of CH$_3$I over Africa (regions 17 and 18) during the Trace-A mission.

[31] Concentrations of CH$_3$I in the tropics decrease with altitude up to 2–3 km, representing the base of the trade wind inversion (TWI). Transport to higher altitudes generally involves wet convection. Many of the profiles in Figure 7 show enhanced concentrations in the middle and upper troposphere (UT) associated with deep convection. In most cases the overall vertical gradient in the tropical free troposphere is weak, indicating that convective outflow is distributed over all altitudes [Wang et al., 2000, 2002]. Comparison of the simulated and observed profiles indicates that the model reproduces the gradual decrease with altitude in the lower troposphere (LT), the location of the TWI at 2–3 km altitude, and the lack of mean vertical gradient at higher altitudes. There is a tendency in the model, and to a somewhat lesser degree in the observations, for a “C-shaped” vertical profile with high concentrations in the UT reflecting preferential deep convection [Prather and Jacob, 1997].
Although errors in the simulation of MBL concentrations (section 4.1) complicate the interpretation of the vertical profiles in Figure 7 it is in fact the shapes of these profiles, rather than the absolute concentrations, that provide a diagnostic of vertical transport. We define therefore a marine convection index (MCI) as the ratio of the mean CH$_3$I concentration at 8–12 km (upper troposphere, UT) to the mean concentration at 0–2.5 km (lower troposphere, LT) for a given region. The observed and simulated values of the MCI are shown in Table 5 for the Pacific regions of Figure 7. Table 5 also shows the CH$_3$I lifetime for each region, which falls in a narrow range (4.0–4.8 days).

The observed MCI values in Table 5 range from 0.11 (Easter Island) to 0.40 (Fiji in March–April), representing large differences in deep convective activity. During PEM-Tropics B (March–April) the observed MCI is highest over the tropical western Pacific (Fiji) and decreases from west to east, consistent with the large-scale upwelling and subsidence associated with the Walker circulation. During PEM-Tropics A (September), the MCI is highest over the central South Pacific (Tahiti). The relatively low MCIs observed over Hawaii (0.20–0.26) reflect the general subsidence due to the North Pacific High. All regions in the equatorial and South Pacific indicate higher MCIs (greater convective mixing) during PEM-Tropics B than PEM-Tropics A, consistent with the wet season timing of PEM-Tropics B [Wang et al., 2002]. The strongest seasonal difference is over Fiji where the MCI decreases from 0.4 in March–April (PEM-Tropics B) to 0.16 in September (PEM-Tropics A). Higher ratios are also found over Hawaii for PEM-Tropics B (0.26) than for PEM-Tropics A (0.20), for reasons that are less clear since the timing of PEM-Tropics B corresponds to the local dry season.

The model reproduces the observed MCI with no significant bias (overall offset of only +11%) for the ensemble of data in Table 5, indicating that the GEOS data provide on average a good simulation of deep marine convective mass fluxes over the Pacific. However, the
Figure 7. Vertical profiles of CH$_3$I concentrations measured from aircraft and averaged over the regions of Figure 6 (region numbers indicated above each plot correspond to those of Figure 6). Horizontal bars are the standard deviations computed from the individual measurements. The corresponding model profiles for the appropriate months are shown as solid black lines. For regions where sources other than oceanic contribute more than 5% to the total CH$_3$I concentrations in the model, the contributions from individual sources are also shown (oceanic (dashed line), biomass burning (dotted line), and rice paddies (dash-dotted line)).
correlation between simulated and observed values of the MCI is weak ($r^2 = 0.15$). The model reproduces qualitatively the spatial gradients and seasonal variations observed over the South Pacific. The model indicates greater deep convection over Hawaii during PEM-Tropics A than during PEM-Tropics B, consistent with the local wet season but contrary to the observations.

5. Summary

[15] We used an oceanic mixed layer model coupled to the GEOS-CHEM 3-D atmospheric transport model to provide a first global simulation of oceanic and atmospheric CH$_3$I. The model was driven by GEOS assimilated meteorological observations from the NASA Data Assimilation Office. Model results were evaluated with an extensive data base of global observations for CH$_3$I in the atmosphere and the oceans. We applied the results to understand the factors controlling the global budget of CH$_3$I and to examine the usefulness of CH$_3$I as a tracer of marine convection in global atmospheric models, complementing the use of $^{222}$Rn as a tracer of continental convection.

[16] The concentration of CH$_3$I(aq) in the oceanic mixed layer was modeled by assuming steady state between photochemical production (dependent on solar radiation and DOC), chemical reaction with Cl$^-$ ions (dependent on temperature), and exchange with the atmosphere (dependent on wind speed and on the local atmospheric concentration of CH$_3$I). This formulation accounts for 40% of the variance in the ensemble of seawater observations. The model predicts high concentrations in the spring-summer hemisphere at midlatitudes where solar radiation is high and seawater temperature is relatively low, but corresponding observations in both the ocean and the air show much lower values than the model. The discrepancy may be possibly due to a missing biological sink for CH$_3$I(aq). Measurements of the biological degradation of CH$_3$I(aq) are needed.

[17] We calculate a gross production of CH$_3$I(aq) in the ocean of 477 Gg yr$^{-1}$ and a net global CH$_3$I source from the ocean to the atmosphere of 214 Gg yr$^{-1}$. The total emission of CH$_3$I from all sources in the model is 304 Gg yr$^{-1}$, including additional contributions from rice paddies (71 Gg yr$^{-1}$), wetlands (7 Gg yr$^{-1}$), biomass burning (9 Gg yr$^{-1}$), and wood fuel (3 Gg yr$^{-1}$). Comparison of model results to observations off the coast of China suggests that the estimate of the source from rice paddies [Redeker et al., 2000] is too high.

[18] We used vertical profiles of CH$_3$I observed from aircraft to test the model simulation of vertical transport over the oceans. Observations in the tropical marine atmosphere indicate a gradual decrease up to the trade wind inversion (TWI) at 2–3 km, a sharp transition across the TWI, and little vertical gradient through the rest of the troposphere reflecting convective outflow at all altitudes. We find that the GEOS meteorological fields reproduce qualitatively these features. We went on to use the ratio of observed CH$_3$I in the upper troposphere (8–12 km) to the lower troposphere (0–2.5 km) as a marine convection index (MCI). The observed MCI over the Pacific ranges from 0.11 for strongly subsiding regions (southeastern subtropical Pacific) to 0.40 for strongly upwelling regions (western equatorial Pacific). The GEOS-CHEM model reproduces the observed MCI values over the Pacific with no significant bias (overall offset of only +11%), implying a good mean simulation of marine convection, but the correlation with the observed MCI for individual regions is weak ($r^2 = 0.15$). The model reproduces qualitatively the spatial gradients and seasonal variations of the MCI observed over the South Pacific, but the seasonal variations are reversed over Christmas Island and Hawaii. We propose that the MCI can provide a useful diagnostic for testing the simulation of marine convection in global atmospheric models.

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