TransCom model simulations of CH and related species: linking transport, surface flux and chemical loss with CH variability in the troposphere and lower stratosphere

The MIT Faculty has made this article openly available. Please share how this access benefits you. Your story matters.

<table>
<thead>
<tr>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>As Published</td>
<td><a href="http://dx.doi.org/10.5194/acp-11-12813-2011">http://dx.doi.org/10.5194/acp-11-12813-2011</a></td>
</tr>
<tr>
<td>Publisher</td>
<td>Copernicus Publications on behalf of the European Geosciences Union</td>
</tr>
<tr>
<td>Version</td>
<td>Final published version</td>
</tr>
<tr>
<td>Accessed</td>
<td>Tue Jan 01 19:51:41 EST 2019</td>
</tr>
<tr>
<td>Citable Link</td>
<td><a href="http://hdl.handle.net/1721.1/70912">http://hdl.handle.net/1721.1/70912</a></td>
</tr>
<tr>
<td>Terms of Use</td>
<td>Creative Commons Attribution</td>
</tr>
<tr>
<td>Detailed Terms</td>
<td><a href="http://creativecommons.org/licenses/by/3.0/">http://creativecommons.org/licenses/by/3.0/</a></td>
</tr>
</tbody>
</table>
TransCom model simulations of CH$_4$ and related species: linking transport, surface flux and chemical loss with CH$_4$ variability in the troposphere and lower stratosphere

P. K. Patra$^{1}$, S. Houweling$^{2,3}$, M. Krol$^{4,3,2}$, P. Bousquet$^5$, D. Belikov$^6$, D. Bergmann$^7$, H. Bian$^8$, P. Cameron-Smith$^7$, M. P. Chipperfield$^9$, K. Corbin$^{10}$, A. Fortems-Chiney$^5$, A. Fraser$^{11}$, E. Gloor$^9$, P. Hess$^{12}$, A. Ito$^{6,1}$, S. R. Kawa$^8$, R. M. Law$^{10}$, Z. Loh$^{10}$, S. Maksyutov$^6$, L. Meng$^{12,*}$, P. I. Palmer$^{11}$, R. G. Prinn$^{13}$, M. Rigby$^{13}$, R. Saito$^1$, and C. Wilson$^9$

$^1$Research Institute for Global Change/JAMSTEC, 3173-25 Show-machi, Yokohama, 2360001, Japan
$^2$SRON Netherlands Institute for Space Research, Sorbonnelaan 2, 3584 CA Utrecht, The Netherlands
$^3$Institute for Marine and Atmospheric Research Utrecht (IMAU), Princetonplein 5, 3584 CC Utrecht, The Netherlands
$^4$Wageningen University and Research Centre, Droevendaalsesteeg 4, 6708 PB Wageningen, The Netherlands
$^5$Universite de Versailles Saint Quentin en Yvelines (UVSQ), GIF sur Yvette, France
$^6$Center for Global Environmental Research, National Institute for Environmental Studies, 16-2 Onogawa, Tsukuba, Ibaraki, 305-8506, Japan
$^7$Atmospheric, Earth, and Energy Division, Lawrence Livermore National Laboratory, 7000 East Avenue, Livermore, CA 94550, USA
$^8$Goddard Earth Sciences and Technology Center, NASA Goddard Space Flight Center, Code 613.3, Greenbelt, MD 20771, USA
$^9$Institute for Climate and Atmospheric Science, School of Earth and Environment, University of Leeds, Leeds, LS2 9JT, UK
$^{10}$Centre for Australian Weather and Climate Research, CSIRO Marine and Atmospheric Research, 107–121 Station St., Aspendale, VIC 3195, Australia
$^{11}$School of Geosciences, University of Edinburgh, King’s Buildings, West Mains Road, Edinburgh, EH9 3JN, UK
$^{12}$Cornell University, 2140 Snee Hall, Ithaca, NY 14850, USA
$^{13}$Center for Global Change Science, Building 54, Massachusetts Institute of Technology, Cambridge, MA, 02139, USA

*now at: Department of Geography and Environmental Studies Program, Western Michigan University, Kalamazoo, MI 49008, USA

Received: 21 June 2011 – Published in Atmos. Chem. Phys. Discuss.: 1 July 2011
Revised: 1 December 2011 – Accepted: 2 December 2011 – Published: 19 December 2011

Abstract. A chemistry-transport model (CTM) intercomparison experiment (TransCom-CH$_4$) has been designed to investigate the roles of surface emissions, transport and chemical loss in simulating the global methane distribution. Model simulations were conducted using twelve models and four model variants and results were archived for the period of 1990–2007. All but one model transports were driven by reanalysis products from 3 different meteorological agencies. The transport and removal of CH$_4$ in six different emission scenarios were simulated, with net global emissions of 513±9 and 514±14 Tg CH$_4$ yr$^{-1}$ for the 1990s and 2000s, respectively. Additionally, sulfur hexafluoride (SF$_6$) was simulated to check the interhemispheric transport, radon ($^{222}$Rn) to check the subgrid scale transport, and methyl chloroform (CH$_3$CCl$_3$) to check the chemical removal by the tropospheric hydroxyl radical (OH). The results are compared to monthly or annual mean time series of CH$_4$, SF$_6$ and CH$_3$CCl$_3$ measurements from 8 selected background sites, and to satellite observations of CH$_4$ in the upper troposphere and stratosphere. Most models adequately capture the vertical gradients in the stratosphere, the average long-term trends, seasonal cycles, interannual variations (IAVs) and interhemispheric (IH) gradients at the surface sites for SF$_6$, CH$_3$CCl$_3$ and CH$_4$. The vertical gradients of all tracers between the surface and the upper troposphere...
are consistent within the models, revealing vertical transport differences between models. An average IH exchange time of 1.39 ± 0.18 yr is derived from SF\textsubscript{6} time series. Sensitivity simulations suggest that the estimated trends in exchange time, over the period of 1996–2007, are caused by a change of SF\textsubscript{6} emissions towards the tropics. Using six sets of emission scenarios, we show that the decadal average CH\textsubscript{4} growth rate likely reached equilibrium in the early 2000s due to the flattening of anthropogenic emission growth since the late 1990s. Up to 60 % of the IAVs in the observed CH\textsubscript{4} concentrations can be explained by accounting for the IAVs in emissions, from biomass burning and wetlands, as well as meteorology in the forward models. The modeled CH\textsubscript{4} budget is shown to depend strongly on the troposphere-stratosphere exchange rate and thus on the model’s vertical grid structure and circulation in the lower stratosphere. The 15-model median CH\textsubscript{4} and CH\textsubscript{3}CCl\textsubscript{3} atmospheric lifetimes are estimated to be 9.99 ± 0.08 and 4.61 ± 0.13 yr, respectively, with little IAV due to transport and temperature.

1 Introduction

The variability of atmospheric CH\textsubscript{4} depends on the spatio-temporal variations of the surface fluxes, atmospheric transport, and destruction due to OH, Cl and O\textsuperscript{1}D chemistry. In recent years, measurements of CH\textsubscript{4} and related species are being conducted at an increasingly large number of sites at hourly or daily time intervals and with high instrumental precision (Rasmussen and Khalil, 1984; Aoki et al., 1992; Dlugokencky et al., 1998; Cunnold et al., 2002; WDCGG, 2010, for a complete list of observational programs). Satellite CH\textsubscript{4} observations from SCanning Imaging Absorption Spectrometer for Atmospheric Cartography (SCIAMACHY), Atmospheric Infrared Sounder (AIRS), and Greenhouse Gases Observing SATellite (GOSAT) are also becoming available, albeit at a lower precision (Frankenberg et al., 2008; Xiong et al., 2008; Yoshida et al., 2011). Significant developments in understanding the distributions, trends and interannual variations of CH\textsubscript{4} emissions and sinks have been achieved in the past two decades through forward modeling (e.g. Fung et al., 1991; Gupta et al., 1996; Houweling et al., 2000; Dentener et al., 2003; Wang et al., 2004; Patra et al., 2009b). Inverse model results show the ability of the models to reproduce the observed atmospheric CH\textsubscript{4} trends and variabilities within the uncertainty of the processes involved (Hein et al., 1997; Houweling et al., 1999; Mikaloff-Fletcher et al., 2004; Chen and Prinn, 2006; Bousquet et al., 2006; Bergamaschi et al., 2009). However, further improvements (reduction in the posterior emission uncertainty) of inverse modeling results depend on a better quantification of (the errors in) the prior emissions and sinks, and on error reductions in forward model transport. Presently, inverse estimates of global CH\textsubscript{4} emissions range between 500 to 600 Tg yr\textsuperscript{−1}, depending on the transport properties and the chemical loss parameterization in the forward models. Bottom-up estimations of individual flux components vary by even greater percentages (e.g. Matthews and Fung, 1987; Yan et al., 2009).

The performance of atmospheric transport models has been investigated within the TransCom project since the early 1990s for the non-reactive tropospheric species, such as sulfur hexafluoride (SF\textsubscript{6}) and carbon dioxide (CO\textsubscript{2}) (Law et al., 1996, 2008; Denning et al., 1999). Convective parameterizations in CTMs have been tested through simulation of Radon (\textsuperscript{222}Rn), which has a radioactive decay half-life of 3.8 days (e.g. Jacob et al., 1997). The full chemistry model simulations of reactive species with a focus on ozone (O\textsubscript{3}) chemistry have also been tested using multiple CTMs, where CH\textsubscript{4} is treated as a tracer with a prescribed concentration evolution (Stevenson et al., 2006, and references therein). Note that most, if not all, full chemistry models do not treat CH\textsubscript{4} in an interactive manner, because of its long lifetime. The CH\textsubscript{4} lifetime ranged from 6.3 to 12.5 yr due to large range of simulated OH concentrations in the participating models (Stevenson et al., 2006). A more conservative estimate of CH\textsubscript{4} lifetime is required for calculating the global warming potential (GWP) for CH\textsubscript{4} and its impact on climate change, or developing effective emission mitigation policies. According to Shindell et al. (2009) the 100-yr integrated GWP of CH\textsubscript{4} is sensitive to changes in oxidant aerosol precursor emissions and to OH-feedbacks of CH\textsubscript{4} emissions itself. Proper understanding of the CH\textsubscript{4} budget is crucial for these assessments.

The aim of the TransCom-CH\textsubscript{4} experiment is to quantify the role of transport, emission distribution and chemical loss in simulating the interhemispheric (IH) gradient, seasonal cycle, synoptic variation and diurnal cycle of CH\textsubscript{4}. Only the first two, IH gradient and seasonal cycles, are discussed in this paper. The dependence of the CH\textsubscript{4} budget on model vertical transport in the stratosphere is also analysed. We setup a long simulation period (1988–2007, including two years of spin-up) for the following reasons: (1) in the 1990s and 2000s methane growth rates have fluctuated between 15 ppb yr\textsuperscript{−1} to −5 ppb yr\textsuperscript{−1} (Dlugokencky et al., 1998; Simpson et al., 2006; Rigby et al., 2008), and (2) we would like to obtain a better understanding of the role of emissions (using a set of six CH\textsubscript{4} emissions scenarios), chemical loss, and transport model characteristics, such as the stratosphere-troposphere exchange (STE) and the IH exchange time on CH\textsubscript{4} concentration variations in the troposphere. The proposed 18-yr simulation period allows a proper quantification of the removal fluxes in the troposphere and stratosphere and of the influence of transport processes on these removal rates. Since the previous TransCom intercomparison experiments spanned only a few years, the 18 yr of SF\textsubscript{6} simulation allows us to track the interannual variability (IAV) in the IH exchange time for the first time. We also discuss the dependence of CH\textsubscript{4} and CH\textsubscript{3}CCl\textsubscript{3} lifetimes on the model grid structure and transport, as well as the transport and temperature as drivers of IAVs in lifetimes.
In Sect. 2, we describe the experimental protocol, followed by the key information on the participating models and analysis methodology. We focus this analysis (Sect. 3) on the comparison of model results with atmospheric observations of SF\textsubscript{6}, CH\textsubscript{3}CCl\textsubscript{3} and CH\textsubscript{4} at 8 surface sites and the salient differences in model properties. An attempt is also made to understand possible implications of (1) inert tracer (SF\textsubscript{6}) transport or short-lived radioactive tracer (\textsuperscript{222}Rn) transport on the lifetimes and distributions of chemically active species (CH\textsubscript{3}CCl\textsubscript{3}, CH\textsubscript{4}), and (2) the effect of the OH abundance, as constrained by CH\textsubscript{3}CCl\textsubscript{3}, on CH\textsubscript{4}. Simulations of SF\textsubscript{6}, \textsuperscript{222}Rn, CH\textsubscript{3}CCl\textsubscript{3} and six CH\textsubscript{4} emission scenarios are commonly referred to as model tracers. Scope for further analysis using the TransCom-CH\textsubscript{4} database and conclusions are given in Sects. 4 and 5, respectively.

2 Models, measurements and methods

Previous TransCom experiments focused on chemically non-reactive species (SF\textsubscript{6}, CO\textsubscript{2}, \textsuperscript{222}Rn). A CH\textsubscript{4} intercomparison requires the introduction of atmospheric chemistry. Additionally, the sources and atmospheric lifetime of CH\textsubscript{4} are distinctly different from CO\textsubscript{2}, which may provide a different view on transport model differences. Detailed documentation of the requested simulation is available in the TransCom-CH\textsubscript{4} protocol (Patra et al., 2010). Chemistry-transport model simulations were requested for the period of 1 January 1990 to 31 December 2007, after a spin-up of 2-yr (1988–1989) using analyzed or atmospheric general circulation model (AGCM) meteorology or a combination of both (referred here as AGCM-nudged). A schematic diagram of TransCom-CH\textsubscript{4} model intercomparison set-up is shown in Fig. 1.

2.1 Photochemical and surface loss processes

The following chemical removal reactions for CH\textsubscript{4} (R1–R3) and CH\textsubscript{3}CCl\textsubscript{3} (R4–R6) are prescribed in the forward simulations.

\[
\text{CH}_4 + \text{OH} \xrightarrow{k_{\text{OH}}} \text{CH}_3 + \text{H}_2\text{O} \quad \left[ k_{\text{OH}} = 2.45 \times 10^{-12} \exp(-1775/T) \right] \quad (R1)
\]

\[
\text{CH}_4 + \text{O}^1\text{D} \xrightarrow{k_{\text{O}^1\text{D}}} \text{Products} \quad \left[ k_{\text{O}^1\text{D}} = 1.5 \times 10^{-10} \right] \quad (R2)
\]

\[
\text{CH}_4 + \text{Cl} \xrightarrow{k_{\text{Cl}}} \text{CH}_3 + \text{HCl} \quad \left[ k_{\text{Cl}} = 7.3 \times 10^{-12} \exp(-1280/T) \right] \quad (R3)
\]
\[
\text{CH}_3\text{CCl}_3 + \text{OH} \rightarrow \text{products} \quad \text{(R4)}
\]

\[
\text{CH}_3\text{CCl}_3 \rightarrow \text{deposition} \rightarrow \text{Oceanic CH}_3\text{CCl}_3 (\text{R6})
\]

The temperature-dependent reaction rates \( k \); units: cm\(^3\) molecule\(^{-1}\) s\(^{-1} \) are taken from the JPL synthesis of chemical kinetics (Sander et al., 2006). The monthly-mean OH fields are provided here for online calculation in the model by combining the semi-empirically calculated tropospheric (Spivakovsky et al., 2000) and 2-dimensional (2-D) model simulated stratospheric distributions. For \( \text{CH}_4 \) reactions with \( \text{Cl} \) and \( \text{O}^1\text{D} \) radicals in the stratosphere, parameterized loss rates \( [k_{\text{O}_1\text{D}} \times \text{O}^1\text{D} + k_{\text{Cl}} \times \text{Cl}] \) are provided, which are based on the Cambridge 2-D model (Velders, 1995).

The tropospheric OH field is reduced by 8\%, an amount that was required to optimize the agreement between the TM5 simulated and observed \( \text{CH}_3\text{CCl}_3 \) decline since 2000 (Huijnen et al., 2010). The model simulations performed here allow us to verify whether observations of \( \text{CH}_3\text{CCl}_3 \) can also be reproduced for a longer simulation period, i.e. 1990–2007, by TM5 and a variety of other models. The supplied OH field has about equal OH abundance in the Northern Hemisphere (NH) and the Southern Hemisphere (SH) (Spivakovsky et al., 2000). Since the NH/SH OH-ratio in full chemistry model simulations varies between 1.1 and 1.5 (Krol et al., unpublished data, 2008, based on the model intercomparison described in Shindell et al., 2006) we encouraged modelers to submit another set of simulations using their preferred OH field, e.g. obtained by a full chemistry version of their model.

For \( \text{CH}_3\text{CCl}_3 \) (MCF), the photolysis rates \( J \) due to solar UV radiation are provided from an Atmospheric general circulation model-based CTM (ACTM; Patra et al., 2009b) and interpolated on each model’s grid. Because the resolution in the stratosphere varies widely between models, it is necessary to scale the stratospheric loss of MCF to a common value. This value is calculated by mass-weighted averaging:

\[
\sum_{i,j,k} J_{\text{CH}_3\text{CCl}_3}(i,j,k) \times M(i,j,k) 
= \sum_{i,j,k} M(i,j,k)
\]

Here, \( M(i,j,k) \) denotes the air mass in gridbox \( (i,j,k) \) from ACTM. Modelers were required to scale their interpolated \( J_{\text{MCF}} \) field to match the \( J_{\text{av, MCF}} \) field of the mass-weighted annual and global mean \( J_{\text{av, MCF}} \) value of 7.959 \( \times 10^{-8} \text{ s}^{-1} \). Similarly, the annual and global mean rate constant for \( \text{CH}_4 \) oxidation due to stratospheric Cl and \( \text{O}^1\text{D} \) combined is rescaled to 2.069 \( \times 10^{-10} \text{ s}^{-1} \).

The monthly deposition velocities \( (\text{deposition}_{\text{CH}_3\text{CCl}_3}) \); units: m s\(^{-1} \) of \( \text{CH}_3\text{CCl}_3 \) to ocean surfaces are provided by Krol et al. (1998; see also Kanakidou et al., 1999). This sink should be applied in the model as:

\[
\text{CH}_3\text{CCl}_3 = \text{deposition}_{\text{CH}_3\text{CCl}_3}(0) \times \exp \left( -\text{deposition}_{\text{CH}_3\text{CCl}_3} \times \frac{1}{dz} \times dt \right)
\]

where, \( dz = \) atmospheric lowest layer depth \( (\text{m}) \), \( dt = \) timestep \( (\text{s}) \), and subscript 0 indicates initial concentration.

Radon decays in the atmosphere with a half-life of 3.8 days, and this decay is calculated in the model at each timestep, following

\[
222\text{Rn} = \left(222\text{Rn}_0\right) \times \exp(-dt \times 2.11 \times 10^{-6})
\]

where \( 222\text{Rn} \) is the radon mixing ratio at all gridpoints. This setup follows the recommendation of World Climate Research Programme (Jacob et al., 1997).

Due to the long timescales of \( \text{CH}_4 \) and \( \text{CH}_3\text{CCl}_3 \) oxidation and vertical transport in the stratosphere (age-of-air \( \sim 5 \text{ yr} \)), several years of spin-up are required to establish realistic \( \text{CH}_4 \) and \( \text{CH}_3\text{CCl}_3 \) vertical profiles throughout the model atmosphere. A set of 3-D initial conditions, prepared following a 10-yr spin-up simulation by ACTM, were made available for 1 January 1988 for \( \text{CH}_4 \), \( \text{SF}_6 \) and \( \text{CH}_3\text{CCl}_3 \). \( \text{CH}_4 \), \( \text{SF}_6 \) and \( \text{CH}_3\text{CCl}_3 \) concentrations at South Pole (SPO) are 1655 ppb, 1.95 ppt and 130 ppt, respectively, for January 1988. Radon will be spun-up quickly due to its half-life of several days. Hence, its initial concentration is set to zero.

2.2 Fluxes

The typical seasonal variations of the six \( \text{CH}_4 \) emission scenarios are shown in Fig. 2a. Annual total emissions time series are depicted in Fig. 2b for \( \text{CH}_4 \), and in Fig. 2c for \( \text{SF}_6 \) and \( \text{CH}_3\text{CCl}_3 \). The following source and sink components of \( \text{CH}_4 \) were considered in the six different scenarios listed in Table 1:

1. Interannually varying anthropogenic emissions (IAV ANT), based on annual mean \( 1^\circ \times 1^\circ \) maps from the Emission Database for Global Atmospheric Research (EDGAR; version 3.2/FT) (Olivier and Berdowski, 2001). The combination of different emission categories and the inter/extra-polation of EDGAR emission maps for the years 1990, 1995, 2000 are described elsewhere (Patra et al., 2009b).

2. Anthropogenic emissions (IAV ANT E4), based on a more advanced EDGAR database (version 4.0) (http://edgar.jrc.ec.europa.eu), where \( 1^\circ \times 1^\circ \) emission maps are available for each year until 2005. The 2005 emissions were also used for the 2006–2008 period.
Table 1. List of tracers simulated in the TransCom-CH$_4$ intercomparison project. See Sect. 2.2 for a description of the CH$_4$ flux components; e.g. cyclostationary natural (CYC NAT), interannually varying anthropogenic (IAV ANT), biomass burning (BB), wetland (WL).

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Description</th>
<th>Time resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH$_4$.CTL</td>
<td>CYC NAT (CYC BB &amp; CYC WL)</td>
<td>Monthly; Partial IAV + IAV ANT EDGAR 3.2</td>
</tr>
<tr>
<td>CH$_4$.CTL.E4</td>
<td>CYC NAT + IAV ANT EDGAR4.0</td>
<td>Monthly; Partial IAV</td>
</tr>
<tr>
<td>CH$_4$.BB</td>
<td>CYC NAT – 0.35 CYC BB + IAV BB</td>
<td>Monthly; Partial IAV + IAV ANT EDGAR 3.2</td>
</tr>
<tr>
<td>CH$_4$.WL.BB</td>
<td>CYC NAT – 0.35 CYC BB + IAV BB – CYC WL + 0.76 IAV WL + IAV ANT EDGAR 3.2</td>
<td>Monthly; Full IAV</td>
</tr>
<tr>
<td>CH$_4$.INV</td>
<td>IPSL/LSCE inversion</td>
<td>Monthly; Full IAV</td>
</tr>
<tr>
<td>CH$_4$.EXTRA*</td>
<td>CYC NAT – 0.35 CYC BB + IAV BB – CYC WL – Rice + IAV WLe (0.69 Wetland + 0.895 Rice) + IAV ANT EDGAR 3.2</td>
<td>Monthly; Full IAV</td>
</tr>
</tbody>
</table>

Other species/tracers

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Description</th>
<th>Time resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>SF$_6$</td>
<td>EDGAR4.0; Global totals modified</td>
<td>Annual; Full IAV</td>
</tr>
<tr>
<td>$^{222}$Rn</td>
<td>1.0 and 0.1 atom m$^{-2}$ s$^{-1}$ over land and ocean, respectively</td>
<td>Annual; No IAV</td>
</tr>
<tr>
<td>CH$_3$CCl$_3$ (MCF)</td>
<td>EDGAR3.2 with trends and distributions modified</td>
<td>Annual; Full IAV</td>
</tr>
</tbody>
</table>

* this scenario is called EXTRA because the VISIT terrestrial ecosystem model (Ito and Inatomi, 2011) fluxes are still under evaluation, but included here since no other bottom-up wetland emission scenario was available with IAV for the full simulation period at the time the intercomparison protocol was released. VISIT is driven by climate variables from the Climate Research Unit time series version 3.0 (CRU TS3.0) dataset (Mitchell and Jones, 2005; updated values) and NCEP/NCAR reanalysis (Kalnay et al., 1996) for the periods of 1988–2005 and 2006–2007, respectively, and CH$_4$ cycling in the inundated areas is modeled using the scheme of Cao et al. (1998, and references therein).

3. Cyclostationary natural emissions (CYC NAT), such as those from all types of natural wetlands, domestic and large-scale biomass burning, and termites, based on the GISS inventory (Matthews and Fung, 1987; Fung et al., 1991), and emissions due to rice paddies taken from Yan et al. (2009). All these emissions are scaled as in Patra et al. (2009b). Though predominantly anthropogenic, emissions from rice cultivation are included in this category because its seasonal cycle is controlled by seasonal rainfall and temperature. The emissions due to oceanic exchange ($\approx$10 Tg CH$_4$ yr$^{-1}$) are distributed over the coastal region (Lambert and Schmidt, 1993; Houweling et al., 1999) and mud volcano emissions are based upon Etiope and Milkov (2004).

4. Wetland emissions with interannual variation (IAV WL) have been derived from the wetland emission module of the ORganizing Carbon and Hydrology in Dynamic EcosystEms (ORCHIDEE) terrestrial ecosystem model (Ringeval et al., 2010). This model uses satellite-derived area of inundation for the period of 1994–2000 (Prigent et al., 2007). The emission is scaled by a multiplication factor of 0.76 to match the wetland emission component in CYC NAT. An average seasonal cycle is used for the rest of the simulation periods (1988–1993 and 2001–2008).

5. Second set of wetland and rice emissions (IAV WLe) is obtained from the Vegetation Integrative Simulator for Trace gases (VISIT) terrestrial ecosystem model (Ito and Inatomi, 2011), which calculates inundated area based on analyzed rainfall, temperature (Mitchell and Jones, 2005). The rice and wetland emissions are scaled by 0.895 and 0.69, respectively, to match with CYC NAT.

6. Biomass burning emissions (IAV BB) are taken from the Global Fire Emission Database (GFED version 2), representing mainly forest and savannah burning (van der Werf et al., 2006). Since this dataset is available only after 1997, an average seasonal cycle is used for the 1988–1996 period. Unlike the wetland emissions, global total IAV BB emissions are lower than those incorporated in CYC NAT. Thus biomass burning emissions in CYC NAT is only partially replaced by IAV BB (see Table 1). This methodology is likely to double count some of the open burning and to underestimate the emissions from closed burning.
Inversion-derived emissions (IAV INV) are obtained by optimizing surface fluxes to reproduce the measured CH$_4$ concentrations using the LMDZ model for the period of 1988–2005 (Bousquet et al., 2006). An average seasonal cycle is repeated for 2006–2008. Concerning the wind field and other meteorology, all models, except ACCESS, used meteorological fields) are given in Table 2 for guidance purposes.

The integrated CH$_4$ emissions for the different combinations of emission fields (scenario 1 to 6 in Table 1) agree within 3 Tg CH$_4$ over 1990–2005 (8675 Tg CH$_4$ for CH$_4_{\text{CTL}}$, which is 542 Tg yr$^{-1}$ on average). Only the IAV INV scenario total emissions are slightly lower (8641 Tg CH$_4$).

Three other species (SF$_6$, $^{222}$Radon and CH$_3$CCl$_3$) are simulated using the following fluxes:

1. Annual mean SF$_6$ emission distributions at $1^\circ \times 1^\circ$ are taken from EDGAR 4.0 (2009) for the period 1988–2005, and the global totals are scaled to Levin et al. (2010). The 2005 distribution is used from 2006 onwards. SF$_6$ emissions increased from 4.77 Tg yr$^{-1}$ in 1990 to 6.79 Tg yr$^{-1}$ in 2007.

2. Radon emissions are constructed based on the surface type in each model grid-cell: 0 poleward of 70°, 8.23 × 10$^{-23}$ mol m$^{-2}$ s$^{-1}$ for 60°–70°, and 1.66 × 10$^{-20}$ and 8.30 × 10$^{-23}$ mol m$^{-2}$ s$^{-1}$ for land and ocean grids, respectively, within 60° S–60° N (Jacob et al., 1997). Radon emission fields were not rescaled to match a global total source, but are expected to produce a global radon source of approximately 2.2 × 10$^{-6}$ mol s$^{-1}$.

3. The annual mean CH$_3$CCl$_3$ emission distribution is taken from EDGAR3.2 and linearly corrected for the global totals following McCulloch and Midgley (2001) for the period 1988–1998. Emissions for 1999 to 2002 are taken as 27.5, 26.0, 17.7, and 16.1 Gg yr$^{-1}$, respectively. After 2002, the regional emission trends follow an exponential decay with a timescale of 5 yr (Krol et al., 2003; updated).

2.3 Participating models and output

Twelve chemistry-transport models and four of their variants (2 at higher horizontal resolution and 2 using different OH, Cl and O(1D) fields) have submitted simulation results for the period 1990–2007 (Table 2). Half of these models (ACTM, CCAM, IMPACT, LMDZ, PCTM, TM5) also participated in the previous TransCom continuous experiment, where they were tested for interhemispheric transport using SF$_6$, and synoptic and diurnal scale variability using continuous CO$_2$ measurements at surface sites (Law et al., 2008; Patra et al., 2008). Six other models (ACCESS, CAM, GEOS-Chem, MOZART, NIES08i, TOMCAT) participated for the first time in a TransCom experiment. The model horizontal resolution varied from $1^\circ \times 1^\circ$ longitude $\times$ latitude to $6^\circ \times 4^\circ$. In the vertical, 19 to 67 levels were employed. Salient features of each model configurations (resolutions, meteorological fields) are given in Table 2 for guidance purpose only, and do not automatically link with model performance as evaluated for various features in this study.

Concerning the wind field and other meteorology, all models, except ACCESS, used meteorological fields from weather forecast models either by interpolation (offline models) or by nudging towards horizontal winds ($U$, $V$) and temperature (online models). Most models generated output as
Table 2. Overview of participating transport models and model variants, and average lifetimes of atmospheric CH$_4$ and CH$_3$CCl$_3$ are given.

<table>
<thead>
<tr>
<th>Sl. No.</th>
<th>Model name$^a$</th>
<th>Institution$^b$</th>
<th>Resolution</th>
<th>Meteorology$^c$</th>
<th>Avg. lifetime (1992–07)$^d$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Horizontal$^c$</td>
<td>Vertical$^d$</td>
<td>CH$_4$-CTL</td>
</tr>
<tr>
<td>1</td>
<td>ACCESS</td>
<td>CSIRO</td>
<td>3.75 × 2.5°</td>
<td>38</td>
<td>AGCM; SST</td>
</tr>
<tr>
<td>2</td>
<td>ACTM</td>
<td>RIGC</td>
<td>~2.8 × 2.8°</td>
<td>67σ</td>
<td>NCEP2; U, V, T; SST</td>
</tr>
<tr>
<td>2a</td>
<td>ACTM_OH$^g$</td>
<td>RIGC</td>
<td>~2.8 × 2.8°</td>
<td>67σ</td>
<td>NCEP2; U, V, T; SST</td>
</tr>
<tr>
<td>3</td>
<td>CAM</td>
<td>CU</td>
<td>2.5 × 1.9°</td>
<td>28σ</td>
<td>NCEP/NCAR</td>
</tr>
<tr>
<td>4</td>
<td>CCAM</td>
<td>RIGC</td>
<td>~220 km</td>
<td>18σ</td>
<td>NCEP; U, V; SST</td>
</tr>
<tr>
<td>5</td>
<td>GEOS-Chem</td>
<td>UoE</td>
<td>2.5 × 2.0°</td>
<td>30/40η</td>
<td>NASA/GSFC/GEOS4/5</td>
</tr>
<tr>
<td>5a</td>
<td>GEOS-Chem_DOH</td>
<td>UoE</td>
<td>2.5 × 2.0°</td>
<td>30/40η</td>
<td>NASA/GSFC/GEOS4/5</td>
</tr>
<tr>
<td>6</td>
<td>IMPACT</td>
<td>LLNL</td>
<td>5.0 × 4.0°</td>
<td>55η</td>
<td>NASA/GSFC/GEOS4</td>
</tr>
<tr>
<td>6a</td>
<td>IMPACT_1 × 1.25</td>
<td>LLNL</td>
<td>1.25 × 1.0°</td>
<td>55η</td>
<td>NASA/GSFC/GEOS4</td>
</tr>
<tr>
<td>7</td>
<td>LMDZ</td>
<td>LSCE</td>
<td>3.75 × 2.5°</td>
<td>19η</td>
<td>ECMWF; U, V; SST</td>
</tr>
<tr>
<td>8</td>
<td>MOZART</td>
<td>MIT</td>
<td>~1.8 × 1.8°</td>
<td>28σ</td>
<td>NCEP/NCAR</td>
</tr>
<tr>
<td>9</td>
<td>NIES08i</td>
<td>NIES</td>
<td>2.5 × 2.5°</td>
<td>32σ–d</td>
<td>JCDAS, ERA-interim-PBL</td>
</tr>
<tr>
<td>10</td>
<td>PCTM</td>
<td>GSFC</td>
<td>1.25 × 1.0°</td>
<td>58η</td>
<td>NASA/GSFC/GEOS5</td>
</tr>
<tr>
<td>11</td>
<td>TM5</td>
<td>SRON</td>
<td>6.0 × 4.0°</td>
<td>25η</td>
<td>ECMWF, ERA-interim</td>
</tr>
<tr>
<td>11a</td>
<td>TM5_1 × 1</td>
<td>SRON</td>
<td>1.0 × 1.0°</td>
<td>25η</td>
<td>ECMWF, ERA-interim</td>
</tr>
<tr>
<td>12</td>
<td>TOMCAT</td>
<td>UoL</td>
<td>~2.8 × 2.8°</td>
<td>60η</td>
<td>ECMWF; ERA-40/interim</td>
</tr>
</tbody>
</table>

$^a$ CTMs driven by AGCM transport are identified in bold (nudging parameters in right-most column), and model variants are shown in italics. The model variants are indicated by post-fixed parameters, following a “$^g$”.

$^b$ CSIRO: Commonwealth Scientific and Industrial Research Organisation, Australia; GSFC: NASA Goddard Space Flight Center, USA; RIGC: Research Institute for Global Change, Japan; CU: Cornell University, USA; LLNL: Lawrence Livermore National Laboratory, USA; LSCE: Laboratoire des Sciences du Climat et de l’Environnement, France; NIES: National Institute for Environmental Studies, Japan; SRON: Netherlands Institute for Space Research; UoE: University of Edinburgh, UK; UoL: University of Leeds, UK.

$^c$ Longitude × latitude or distance or spectral resolution indicated by T (triangular) maximum wave number (T42 and T63 for ~2.8 × 2.8° and ~1.8 × 1.8°, respectively).

$^d$ Terrain-following (height) coordinate system for ACCESS, σ vertical coordinates are pressure divided by surface pressure, η vertical coordinates are a hybrid sigma-pressure coordinate (GEOS-Chem has 30 or 47 layers for 1990–2006 or 2007, respectively), NIES08i has a hybrid sigma-isentropic.

$^e$ The source of meteorology (NCEP2 (AMIP DOE II): Kanamitsu et al., 2002; NCEP: Kalnay et al., 1996; NASA/GSFC/GEOS4/5: Bloom et al., 2005; ECMWF: Uppala et al., 2005; JCDAS: Onogi et al., 2007) and parameters used in nudged AGCMs are given.

$^f$ The averaging period for IMPACT$^{1} × 1.25$ (2002–2007) and TM5$^{1} × 1$ (2003–2007) differ.

$^g$ The tropospheric OH field is taken from CHASER full chemistry model (Sudo et al., 2002) and scaled by ×0.88, and stratospheric OH is taken from AGCM (Takigawa et al., 1999) as discussed in Patra et al. (2009a). SF$_6$ emissions are used from EDGAR4.0 without scaling the global totals to match with Levin et al. (2010).

1-hourly averages, except LMDZ and MOZART, which provided output as 3-hourly averages.

Details of individual transport models can be found in the following references; ACCESS (Corbin and Law, 2011), ACTM and ACTM_OH (Patra et al., 2009a, b), CAM (Gent et al., 2009), CCAM (Law et al., 2006), GEOS-Chem and GEOS-Chem_DOH (Pickett-Heaps et al., 2011; Fraser et al., 2011), IMPACT and IMPACT$^{1} × 1.25$ (Rotman et al., 2004), LMDZ (version 4; Houdrin et al., 2006), MOZART (version 4; Emmons et al., 2010), NIES08i (Belikov et al., 2011), PCTM (Kawa et al., 2004), TM5 and TM5$^{1} × 1$ (Krol et al., 2005), TOMCAT (Chipperfield, 2006).

The TransCom-$\text{CH}_4$ experiment archived model simulations for 18 yrs and 9 tracers. We have sampled model output at 280 surface sites and 115 vertical profile sites (at all model levels within the troposphere) at hourly time intervals, 3-D output at 17 standard pressure levels for monthly-means for 1990–2007, and noon-time daily values for 2001–2007 are also archived.

2.4 Observational data sources and processing

Selected sites from the Advanced Global Atmospheric Gases Experiment (AGAGE; http://agage.eas.gatech.edu) and the NOAA Earth System Research Laboratory, Global Monitoring Division (http://www.esrl.noaa.gov/gmd) networks are used in this study. These sites all have simultaneous measurements of CH$_4$, SF$_6$, and CH$_3$CCl$_3$ covering the 1990s and 2000s (Table 4). Unfortunately, radon measurements are not available for most of these sites. Monthly or annual mean observations have been calculated from continuous (hourly averages) or flask sampling (events) measurements data available from the World Data Center for Greenhouse Gases website (WDCCG, 2010). NOAA flasks are usually sampled under clean air (or baseline) conditions; this is usually onshore flow at coastal sites. The AGAGE continuous records have been flagged to remove local and regional pollution events.

The model outputs are extracted for the corresponding sites and sampling time from the hourly surface data files. All the models sampled concentrations for BRW and CGO stations at the nearest full ocean grid, i.e. BRWOCN and CGOOCN, respectively, to better represent baseline
conditions. However, note that some models, e.g. TM5, interpolate model output to the site locations, and submit identical values at the land and ocean grids for the coastal sites. The results of HBA (75.6° S, 26.5° W, 10 m) site are used as a replacement for the SPO site for PCTM. These selections are made as per the modeler’s advice.

For CH₄, the NOAA-2004 calibration scale (Dlugokencky et al., 2005) agrees within 5 pptb with the AGAGE/Tohoku University calibration scale (Aoki et al., 1992). The calibration scales for CH₃CCl₃ are as per the Scripps Institution of Oceanography (SIO) 2005 (Prinn et al., 2005) for AGAGE and NOAA-2003 (http://www.esrl.noaa.gov/gmd/ccl/scales/CH3CCl3_scale.html) for NOAA sites. Those for SF₆ are based on SIO 2005 (Rigby et al., 2010) and NOAA-2006 (http://www.esrl.noaa.gov/gmd/ccl/sf6_scale.html). The AGAGE and NOAA scales for SF₆ are in excellent agreement (NOAA–AGAGE = 0.02 ± 0.01 ppt, compared to repeatability of the analytical system of 0.04 ppt) (Rigby et al., 2010). The systematic NOAA–AGAGE difference of ~4 ppt in 1992 reduces linearly to around 0 ppt in 2001 for the CH₃CCl₃ concentrations as determined from co-located measurements at CGO, SMO and MHD by both the networks (P. Krummel and T. Arnold, personal communication, 2010).

As a check for the stratospheric CH₄ distribution simulated by the models, a climatology of CH₄ vertical profiles measured by the ACE-FTS (Atmospheric Chemistry Experiment – Fourier Transform Spectrometer) instrument on board the SCISAT-1 satellite in the upper troposphere and stratosphere altitudes has been used (De Mazière et al., 2008). Please note that the data coverage of the ACE-FTS instrument in the tropical region is sparse. The HALOE/UARS (Halogen Occultation Experiment onboard the Upper Atmosphere Research Satellite) (Park et al., 1996) had a denser coverage and measurements of stratospheric CH₄ from this are also used for validating simulated vertical gradients in the tropical stratosphere (100–10 mb).

For this analysis, the models were sampled within 1 to 3 h of the sampling times of the measurements. Time series were constructed of monthly or annual mean samples for verification of model simulated seasonal cycles and interannual variability, respectively. The Pearson’s moment correlation (Press et al., 1986) analysis is performed to evaluate the agreements between the simulated and observed time series for seasonal cycles and interannual variations at the 8 selected sites.

2.5 Calculation of IH gradients, IH exchange time, and atmospheric lifetimes

The IH exchange time (τ_ex) is estimated from the SF₆ annual mean concentration time series and the ratio of emission in the NH (E_n) and SH (E_s) using (Patra et al., 2009a, and references therein):

\[ \tau_{ex} = \frac{\Delta c_{n-s}}{\left( \frac{E_n}{E_s} + 1 \right)} \left[ \frac{E_n}{E_s} \frac{dc_n}{dt} - \frac{dc_s}{dt} \right] \]  

(4)

c_n and c_s are the average concentrations of SH and NH sites, and Δc_{n-s} is IH concentration gradient. Derivation of this equation assumes that Δc_{n-s} is in steady state (Jacob et al., 1987). We used two sites in the NH (BRW, MLO) and SH (CGO, SPO) to estimate the hemispheric average concentrations c_n and c_s, respectively, at yearly time intervals (dt). Because SF₆ measurements at ALT are not available after 2005, BRW is chosen for this analysis. The values c_n - c_s are shown as IH gradients of each species. We have tested that the calculated τ_ex does not depend strongly on the \( \frac{E_n}{E_s} \) ratio.

We also estimated CH₄ and CH₃CCl₃ lifetimes (τ) using the mass balance equation

\[ \frac{dB}{dt} = E - L = E - \frac{B}{\tau} \]  

(5)

B, E and L are the annual total atmospheric burden, net surface emission and photochemical loss, respectively. As approximate estimates of B, average concentrations of CH₄ and CH₃CCl₃ for 8 sites are multiplied by the concentration-to-mass conversion factors of 2.845 Tg CH₄ ppb⁻¹ and 23.689 Gg CH₃CCl₃ ppt⁻¹, respectively. Ideally, vertical distribution properties of each species should be accounted for with appropriate air mass factors for calculating B. However, we note that the average lifetimes over the period of 2000–2007 calculated using ACTM simulated gridded loss rates agree within 0.01 yr for CH₃CCl₃ with the lifetimes calculated using Eq. (5) suggesting that the simplification is acceptable.

3 Results and discussions

3.1 Zonal mean concentrations

Figure 3 compares the latitude-pressure variations of the zonal mean CH₄,CTL scenario in the troposphere and lower stratosphere to representative ACE-FTS measurements (please refer to Figs. S1–S17 for individual model comparison plots of ²²²Rn, SF₆ and CH₄ along the 70° E and 180° E longitudes, and zonal mean CH₃CCl₃ distributions corresponding to the years 1994 and 2005). Generally, all models exhibit similar large-scale features, equator-pole latitudinal gradients, and vertical gradients in the lower stratosphere. Most significantly different, however, is the transition between the troposphere and stratosphere and the heights at which the vertical gradients maximize in the models (the green-to-blue shaded regions). The ACCESS/GEOS-Chem/NIES-08i and CAM/MOZART/CCAM models show the maximum and minimum decrease rates of CH₄ with increasing height, respectively, in the height range of 100–50 mb. CAM, CCAM and MOZART models have only few vertical layers above 100 mb and use the same reanalysis wind fields from NCEP, and the NIES model employs isentropic coordinate system in the stratosphere. The formulation of models is known to affect the simulation of tracer
Fig. 3. Annual and zonal mean latitude-pressure (in mb) cross-sections of CH$_4$ for tropospheric and lower stratospheric altitudes as observed by the ACE-FTS instrument (a; climatology) and as simulated by the models in 2000 (b–q). The black line in (a) shows the climatological tropopause height. The contour lines in (c and d) show CH$_4$ loss rate (units: molecule cm$^{-3}$ s$^{-1}$) as in the ACTM and ACTM-OH, respectively. An offset is added to the concentrations in each panel (given after the model name in ppb) that adjusts the model fields to a common average value of 1770 ppb between 950 mb and 500 mb. Detailed model-to-model comparisons for $^{222}$Rn, SF$_6$, CH$_4$ and CH$_3$CCl$_3$ for two seasons and over two longitudes are available in the Supplement (Figs. S1–S17).
Table 3. Multi-model averages (± 1σ; and between model variability defined by 1σ/average, within parenthesis in %) of simulated SF₆ (ppb), CH₄_CTL (ppb), CH₃CCl₃ (ppt) and ²²²Rn (× 10⁻²¹) gradients in the troposphere and UT/LS region for three broad latitude bands, namely, the SH midlatitude, tropics and NH midlatitude.

<table>
<thead>
<tr>
<th>Species</th>
<th>SH (60–30° S)</th>
<th>Tropics (15° S–15° N)</th>
<th>NH (30–60° N)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Tropospheric gradients</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Difference between:</td>
<td>850–400 mb</td>
<td>850–200 mb</td>
<td>850–400 mb</td>
</tr>
<tr>
<td>SF₆</td>
<td>-0.026 ± 0.005 (18%)</td>
<td>0.044 ± 0.017 (39%)</td>
<td>0.128 ± 0.018 (14%)</td>
</tr>
<tr>
<td>CH₄</td>
<td>-9.91 ± 2.57 (26%)</td>
<td>17.18 ± 7.69 (45%)</td>
<td>44.65 ± 5.91 (13%)</td>
</tr>
<tr>
<td>CH₃CCl₃</td>
<td>0.473 ± 0.145 (31%)</td>
<td>0.295 ± 0.182 (62%)</td>
<td>1.01 ± 0.25 (25%)</td>
</tr>
<tr>
<td>²²²Rn</td>
<td>0.984 ± 0.527 (54%)</td>
<td>8.66 ± 2.73 (32%)</td>
<td>23.90 ± 6.73 (28%)</td>
</tr>
<tr>
<td></td>
<td>UT/LS gradients</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Difference between:</td>
<td>200–100 mb</td>
<td>100–50 mb</td>
<td>200–100 mb</td>
</tr>
<tr>
<td>SF₆</td>
<td>0.158 ± 0.074 (47%)</td>
<td>0.264 ± 0.148 (56%)</td>
<td>0.229 ± 0.140 (61%)</td>
</tr>
<tr>
<td>CH₄</td>
<td>84.45 ± 35.89 (43%)</td>
<td>124.04 ± 65.07 (52%)</td>
<td>128.77 ± 64.96 (50%)</td>
</tr>
<tr>
<td>CH₃CCl₃</td>
<td>7.86 ± 1.37 (17%)</td>
<td>16.59 ± 4.48 (27%)</td>
<td>9.79 ± 3.75 (38%)</td>
</tr>
<tr>
<td>²²²Rn</td>
<td>1.084 ± 0.708 (65%)</td>
<td>0.415 ± 0.203 (49%)</td>
<td>0.353 ± 0.445 (126%)</td>
</tr>
</tbody>
</table>

Table 4. Average (across all models) correlation coefficient (r) between model simulated and observed seasonal cycles and interannual variation of CH₄, SF₆ and CH₃CCl₃.

<table>
<thead>
<tr>
<th>Tracer/Site*</th>
<th>ALT</th>
<th>BRW</th>
<th>MHD</th>
<th>MLO</th>
<th>RPB</th>
<th>SMO</th>
<th>CGO</th>
<th>SPO</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH₄_CTL</td>
<td><strong>0.16 ± 0.3</strong></td>
<td>-0.09 ± 0.2</td>
<td>0.73 ± 0.1</td>
<td>0.63 ± 0.1</td>
<td>0.86 ± 0.1</td>
<td>0.74 ± 0.2</td>
<td>0.98 ± 0.2</td>
<td>0.96 ± 0.1</td>
</tr>
<tr>
<td>CH₄_CTL E4</td>
<td><strong>0.21 ± 0.3</strong></td>
<td>-0.10 ± 0.3</td>
<td>0.72 ± 0.1</td>
<td>0.69 ± 0.1</td>
<td>0.88 ± 0.1</td>
<td>0.71 ± 0.2</td>
<td>0.97 ± 0.2</td>
<td>0.96 ± 0.1</td>
</tr>
<tr>
<td>CH₄_BB</td>
<td><strong>0.10 ± 0.3</strong></td>
<td>-0.11 ± 0.2</td>
<td>0.68 ± 0.1</td>
<td>0.66 ± 0.1</td>
<td>0.86 ± 0.1</td>
<td>0.72 ± 0.2</td>
<td>0.95 ± 0.1</td>
<td>0.96 ± 0.1</td>
</tr>
<tr>
<td>CH₃CCl₃</td>
<td><strong>0.38 ± 0.3</strong></td>
<td>0.48 ± 0.2</td>
<td>0.42 ± 0.2</td>
<td>0.70 ± 0.1</td>
<td>0.86 ± 0.1</td>
<td>0.64 ± 0.3</td>
<td>0.94 ± 0.1</td>
<td>0.95 ± 0.1</td>
</tr>
<tr>
<td>CH₄_INV</td>
<td>0.88 ± 0.1</td>
<td>0.63 ± 0.2</td>
<td>0.92 ± 0.1</td>
<td>0.74 ± 0.1</td>
<td>0.88 ± 0.1</td>
<td>0.61 ± 0.2</td>
<td>0.95 ± 0.2</td>
<td>0.92 ± 0.1</td>
</tr>
<tr>
<td>CH₃EXTRA</td>
<td>0.84 ± 0.1</td>
<td>0.37 ± 0.2</td>
<td>0.79 ± 0.1</td>
<td>0.50 ± 0.2</td>
<td>0.80 ± 0.1</td>
<td>0.69 ± 0.3</td>
<td>0.93 ± 0.1</td>
<td>0.95 ± 0.1</td>
</tr>
<tr>
<td>SF₆</td>
<td>0.48 ± 0.2</td>
<td><strong>0.23 ± 0.2</strong></td>
<td>0.68 ± 0.1</td>
<td><strong>-0.06 ± 0.1</strong></td>
<td><strong>0.13 ± 0.2</strong></td>
<td>0.70 ± 0.4</td>
<td>0.43 ± 0.2</td>
<td>0.50 ± 0.2</td>
</tr>
<tr>
<td>CH₃CCl₃</td>
<td>0.78 ± 0.1</td>
<td>0.93 ± 0.2</td>
<td>0.87 ± 0.1</td>
<td>0.91 ± 0.1</td>
<td>0.96 ± 0.2</td>
<td>0.89 ± 0.1</td>
<td>0.88 ± 0.1</td>
<td><strong>0.81 ± 0.1</strong></td>
</tr>
</tbody>
</table>

* ALT (Alert, Canada; 62° W, 82° N, 210 m), BRW (Point Barrow, Alaska, USA; 157° W, 71° N, 11 m), MLO (Mauna Loa Observatory, Hawaii, USA; 156° W, 20° N, 3397 m) and SPO (South Pole Observatory, Antarctica; 25° W, 90° S, 2810 m) are managed under the NOAA cooperative network by the Global Monitoring Division, Earth System Research Laboratory (GMD/ESRL) (Dlugokencky et al., 1998; Butler et al., 2004), and MHD (Mace Head, Ireland; 10° W, 53° N, 25 m), RPB (Ragged Point, Barbados; 59° W, 13° N, 45 m), SMO (Samoa, USA; 171° W, 14° S, 42 m) and CGO (Cape Grim, Australia; 145° E, 41° S, 94 m) sites are operated under the AGAGE network by the Massachusetts Institute of Technology (MIT, USA), Scripps Institution of Oceanography, University of California, San Diego (SIO/UCSD, USA), Commonwealth Scientific and Industrial Research Organization (CSIRO, Australia), University of Bristol, UK and Georgia Institute of Technology, USA (Cunnold et al., 2002; Prinn et al., 2000).

Gradients in the upper troposphere/lower stratosphere region. A version of the TOMCAT/SLIMCAT model, which uses isentropic coordinates in the stratosphere, produces stronger tracer gradients and a more realistic Brewer-Dobson circulation than the p-coordinate version used here (e.g. Hosaini et al., 2010, and references therein). The “tropical pipe” (Plumb, 1996) along the upward transport branch of the Brewer-Dobson circulation appears more “leaky” in the
models than in the limb-viewing remote sensing observations by HALOE and ACE-FTS. As a result, the simulated concentration isopleths appear flatter compared to the observations with increasing latitudes in both hemispheres. The CH$_4$ meridional gradients between tropics ($\pm 10^\circ$ N) and northern extratropics ($20^\circ$ N–$30^\circ$ N) are calculated to be 140 $\pm$ 72 ppb for the models and 225 ppb for HALOE in the height range of 70–30 mb.

In the troposphere, vertical transport of the NH emission varies between the models, most prominently in the tropical region, where deep cumulus convection is prevalent. Table 3 shows mean vertical gradients and between model variabilities in three broad latitude ranges. Model to model differences are much less distinct at the mid and high latitudes than in the tropics for SF$_6$, CH$_4$ and CH$_3$CCl$_3$. However, two main categories of models can be identified based on the density of CH$_4$ isopleths in the height range between 350 and 200 mb. The position and concentration gradient across the tropopause differs considerably among models. These model features are also present in the SF$_6$ simulations suggesting the predominant role of transport in the simulation of the CH$_4$ vertical distributions. The penetrative mass flux due to deep cumulus convection in tropical latitudes is strongest in ACTM and GEOS-Chem (low $^{222}$Rn concentration difference of $\sim 5.63 \times 10^{-21}$ between 850 and 200 mb), and relatively weaker in NIES08i, PCTM and TOMCAT (high $^{222}$Rn
concentration difference > 12 × 10^{-21}). This is clearer from the simulated $^{222}\text{Rn}$ distributions over the South Asian monsoon region at 70° E during boreal summer (Figs. S2 and S4). It has been shown in Patra et al. (2009a), based on ACTM simulations with and without cumulus parameterization, that deep cumulus convective transport is the main cause for rapid vertical transport of tracers to the upper troposphere (seen as higher $^{222}\text{Rn}$ concentrations compared to the lower troposphere). Feng et al. (2011) showed that the online convection scheme used in the TOMCAT runs for this study underestimated the ECMWF archived convective mass fluxes especially in terms of the altitude extent of deep convection in the tropics. The higher horizontal resolution versions of both IMPACT and TM5 resulted in higher $^{222}\text{Rn}$ concentrations in the middle-upper troposphere (i.e. smaller difference between 850 mb and 400/200 mb) compared to their respective lower resolution simulations. This is suggesting that some convective processes are being resolved in higher resolution models that are not present in the lower resolution models.

3.2 Model-observation comparison of CH$_4$, SF$_6$ and CH$_3$CCl$_3$ annual means: 1990–2007

Figure 4 shows the time series of annual mean concentrations differences between simulated and observed CH$_4$, SF$_6$ and CH$_3$CCl$_3$ at two selected sites (MLO and CGO). First SF$_6$ is considered, which has no chemical loss (middle row). Typical model behaviour is similar at all 8 sites (not shown). For most models, the simulated concentrations diverge from the measurements by 0.2 ppt in 1995, after which differences remain at that level. The offsets between models can be explained by initial values assumed by each models. Only ACCESS shows increasing differences in time until 2006. The ACTM_OH case, which uses EDGAR4.0 emissions without scaling between 1988–2005, and 2005 emission for 2006 and 2007 and produces a slower increase in the model concentration compared to observations after 2000. This suggests that the global total emissions estimated by Levin et al. (2010) and later confirmed by Rigby et al. (2010) are adequate also for independent state-of-the-art transport models.

The CH$_3$CCl$_3$ simulations are of an intermediate level of uncertainty, as this species is emitted to the atmosphere by a relatively well quantified industrial use as a solvent. However, uncertainties remain for its loss by reaction with OH in the troposphere, photolysis in the stratosphere and the rate of STE. Until 1990, CH$_3$CCl$_3$ emissions were increasing, followed by near-exponential decrease due to stringent restriction of its production/use by the Montreal Protocol (WMO/SAOD, 2003). The lifetimes of CH$_3$CCl$_3$ due to photochemical removal is much longer in the stratosphere (~28.6 yr) than in the troposphere (~5.8 yr) (estimates from ACTM simulated loss rates at model grids; similar estimates of lifetimes using TM5,1×1 are 37.6 and 5.8 yr, respectively). Thus the troposphere to stratosphere transport plays a minor role in the global total budget of CH$_3$CCl$_3$ after the late 1990s, because the concentration gradients across the tropopause reduced to less than 10 ppt (ref. Table 3). Although these lifetimes in the stratosphere and troposphere are within the range of independent estimates, 38$^{+15}_{-11}$ and 5$^{+0.5}_{-0.4}$ yr, respectively (Prinn et al., 2005), both the mean values are lower for ACTM. Despite the fact that the models were initialized using the same concentration at 1 January 1988, significant differences in the simulated concentrations are found already for 1990, after two years of simulation. The established model-measurement differences in 1990 persist until the end of the simulations (LMDZ CH$_3$CCl$_3$ being the only outlier) even though the CH$_3$CCl$_3$ concentrations become very small towards 2007. Given the CH$_3$CCl$_3$ lifetime of less than 5 yr, these differences are caused by differences in transport and removal rather than the initialization.

CH$_4$ is the most complicated species considered in the TransCom-CH$_4$ experiment, because the CH$_4$ surface emissions and effect of STE are less certain than for CH$_3$CCl$_3$. Generally, the models that simulate lower CH$_3$CCl$_3$ concentrations compared to the multi-model mean, also yield lower CH$_4$ concentrations (such as MOZART, CCAM). For example, the high and low-resolution TM5 simulations show the highest concentrations of both CH$_4$ and CH$_3$CCl$_3$ at MLO. However, it is interesting to note that GEOS-Chem (with TransCom OH) calculates the lowest CH$_4$ concentrations among all models, while the simulated CH$_3$CCl$_3$ levels are not distinctly different. LMDZ is among the models that most strongly underestimate the observed level of CH$_3$CCl$_3$, whereas the opposite is true for CH$_4$ (Fig. 4). These contrasting behaviors among models for various CH$_4$ and CH$_3$CCl$_3$ simulations clearly suggest that CH$_4$ loss due to the reaction with OH in the troposphere is not the only control on the CH$_4$ budget differences between models and that other factors such as transport differences also play a role (details in Sects. 3.3 and 3.4).

3.2.1 Seasonal cycles

Figure 5 shows model to measurement comparisons of the seasonal cycles of CH$_4$, SF$_6$ and CH$_3$CCl$_3$ at three selected sites (MLO, SMO and CGO) for the period 2002–2003. These sites have been selected because they are at large distances from the continental emissions for each of these species. To highlight differences in seasonality, approximate linear trends and offsets corresponding to the period 2002–2003 have been subtracted from the monthly-mean values. All models capture the salient features in the seasonal cycles at very high statistical significance (ref. Table 4 for correlation coefficients at 8 sites), except for SF$_6$ at MLO, where the measurements show unusual fluctuations and a large data gap during 2002. Even for the years with dense data coverage (2005–2006), low average correlation coefficients (r) of 0.3 for the SF$_6$ seasonal cycles at MLO are obtained due to a very small seasonal cycle of less than 0.04 ppt. In contrast,
the clear seasonalities (amplitude >0.04 ppt) at ALT, MHD, SMO, CGO and SPO are well reproduced by the models \((r > 0.4); \text{ significant at } P = 0.05 \text{ in two tailed Student’s } t\)-test for 24 data points). The fact that the models are able to reproduce the observed seasonal cycles indicates that even though the signals are weak, they nevertheless provide useful information for model validation. All models reproduce the CH$_3$CCl$_3$ seasonal cycle fairly well at all 8 sites, both in phase and amplitude \((r > 0.8)\).

For CH$_4$, the influence of the surface flux on the simulated seasonal cycles can be studied using the 6 different CH$_4$ scenarios. The corresponding correlation coefficients, listed in Table 4, suggest that the CH$_4$ seasonal cycle depends strongly on the implemented wetland and biomass burning fluxes. The CH$_4$INV simulations consistently produce higher correlation coefficients at all the NH sites (ALT, BRW, MHD, MLO and RPB), which was expected because the atmospheric-CH$_4$ inversion used data from these sites for flux optimization. CH$_4$EXTRA results are next best in comparison with measured seasonal cycles for the NH high latitude sites. For any given tracer, the correlation coefficients are highest at remote SH sites, CGO and SPO, compared to all other sites. The use of ACTM$_{OH}$ and GEOS-Chem$_{DOH}$ or the higher horizontal resolution versions of IMPACT$_{1 \times 1.25}$ and TM5$_{1 \times 1}$ do not always improve the agreement between model and observations compared to the default implementation. These results suggest a need for improving our understanding of the CH$_4$ flux seasonality in the Northern Hemisphere land regions, noting that the OH loss is realistically represented as seen in the simulated
Fig. 6. Comparisons of observed and simulated annual mean growth rates of CH$_4$ in NH (top row; ALT, BRW and MHD average), tropics (middle row; MLO RPB and SMO average) and SH (bottom row; CGO and SPO average) for two selected fluxes, CTL (left column) and EXTRA (right column). Growth rate variabilities corresponding 4 other CH$_4$ scenarios are given in Figs. S20 and S21.

CH$_3$CCl$_3$ seasonal cycles. Because ACCESS (blue line) was not run with analyzed winds and temperature, the simulated seasonal cycles are not as good as other models for CH$_4$ and SF$_6$ highlighting the role of meteorology in simulating tracer concentrations. The role of meteorology is less pronounced for CH$_3$CCl$_3$ because the emissions are weak in the 2000s. A more detailed analysis of seasonality at a larger number of sites will be conducted in a future study.

3.2.2 Interannual variability (IAV)

We calculated growth rates for all tracers as the difference between annual mean concentrations for two adjacent years. The growth rate at January 2000 is shown as the difference between 1999 and 2000 mean concentrations. The simulated and observed SF$_6$ growth rates (not shown) decreased from $\sim$0.25 ppt yr$^{-1}$ in 1997 to $\sim$0.2 ppt yr$^{-1}$ in 2000. Afterwards, the growth rate steadily increased to $\sim$0.25 ppt yr$^{-1}$ in 2006 (please refer to Table 4 for correlation coefficients; $r \sim 0.7$ for 5 sites). The length of time series considered for the correlation calculation of the IAVs is 1990–2007 for CH$_4$. 
BB, only CTL lifetime (9.82 yr) is OH version is due to smaller SF.

However, BB), the BB) improves the IA V model-observation simulations capture the observed reduction in the EXTRA displays a large positive anomaly on top of 2011BB, which has 7 yr of IA V, 1994–2000)

in CH 

higher air temperatures (Reaction R1), resulting in faster re-

reaction coefficients for the growth rate IAVs in Table 4; see also Fig. S20). This indicates that forward simulations using multiple forward transport models are useful for the verification of emission inventories.

3.3 Interhemispheric gradients and exchange times

Figure 7 shows the concentration gradients between two NH sites and two SH sites obtained using annual mean observed and modeled time series. All models except for the TM5s, simulate the observed SF 

emissions are of purely anthropogenic origin, their production, consumption and release to the atmosphere vary relatively smoothly in comparison to the natural components of CH 

emissions. The decreasing growth rate in the 1990s, near zero growth rates in early 2000s and the reappearance of positive CH 

growth in the late 2000s have drawn considerable interests for developing emission inventories. For example, Lamarque et al. (2010) suggested a decrease of CH 

emissions by about 40 Tg CH 

yr 

−1 during 2001–2007, synchronized with the Chinese economic growth, produces inconsistencies between observed and simulated growth rates during 2003–2007 (thus the lowest correlation coefficients for the growth rate IAVs in Table 4; see also Fig. S20). This indicates that forward simulations using multiple forward transport models are useful for the verification of emission inventories.

As seen from Table 4, inclusion of biomass burning emission IAV (CH 

BB) improves the IAV model-observation agreement at all sites compared to CH 

CTL. However, when wetland emissions are included (CH 

WL-BB), the correlations tend to deteriorate. Compared to CH 

BB, only CH 

EXTRA produces better model-observation agreement for growth rates (Fig. 6, right panels). The wetland CH 

emission simulated by the VISIT ecosystem model included in CH 

EXTRA displays a large positive anomaly on top of the emissions from biomass burning during 1997/1998. Combined, these emissions compensate for the extra CH 

loss due to higher air temperatures. As a consequence, these emissions result in an excellent agreement with the growth rates observed at the SH and NH sites (except that the simulated tropical signal is a bit too strong in 1998). The IAV of the multi-model average CH 

growth rate at BRW did not correlate significantly with the observed IAV because this site is located close to the Alaskan wetland region, and the
Fig. 7. Interhemispheric gradients (IHGs) for CH$_4$-CTL, SF$_6$ and CH$_3$CCl$_3$ concentrations between the NH (BRW, MLO) and SH (CGO, SPO) sites. The values at CGO (AGAGE) are adjusted to NOAA scales by adding an offset of 0.02 ppt for SF$_6$, and multiplied by 1.0003 and 1.0333 for CH$_4$ and MCF, respectively (see text in Sect. 2.4 for further details). Please note that adjustment of the AGAGE data to NOAA scale is made just for convenience. These 4 sites are chosen here because their data coverage is most complete during 1990–2007. Haley Bay (75.58° S, 26.5° W, 10 m) site is chosen for PCTM due to no SPO data in all files. Inset shows expanded y-axis view of MCF for the 2000–2007 period.

protocol (EDGAR4.0/Levin), (2) the EDGAR4.0 emission distribution corresponding to 2000, but global totals scaled to Levin et al. (2010) (EDGAR$_{Y2000}$/Levin), and (3) constant emissions from EDGAR4.0 for the year 2000 (EDGAR$_{Y2000}$). Figure 9a shows the trends in SF$_6$ emissions within three broad latitude bands. The emissions in the NH mid-high latitudes (circles) remained fairly constant (within 14%) in EDGAR4.0/Levin scenario, while the emissions from the NH tropical latitudes (triangles) increased by more than 100% during 1996 to 2007. The estimated $\tau_{ex}$
using three emission scenarios are shown in Fig. 9b. These results clearly suggest that the decrease in τex is caused by the trends in the SF6 emission distribution rather than inter-hemispheric transport. The decrease in τex becomes an order of magnitude smaller as well as statistically insignificant (r = −0.34, N = 12) when SF6 emissions were kept constant (EDGAR_Y2000). The decrease in τex remained significant for scenario EDGAR_Y2000/Levin (red square/line). This is probably because the SF6 IH gradients do not reach steady state, since the emissions decreased during 1995–1998 and increase thereafter till 2008 (Fig. 2c), even though the δex ratio was fixed at 36.9, corresponding to the year 2000. Our results indicate that the time evolution of the SF6 emissions should be introduced at a higher spatial resolution (than the presently used 2-box model), when calculating τex.

All three long-lived species show a similar relationship for the IH gradient: the model that produces a larger (smaller) SF6 IH gradient generally also produces a larger (smaller) IH gradients for CH3CCl3 and CH4 in comparison with the observations (Fig. 10). The intriguing exception is MOZART, which exhibits an excellent match for SF6 IH gradient, but produces one of the largest CH4 IH gradients (127 ppb compared to an observed value of 101 ppb) and one of the smallest CH3CCl3 gradients (0.13 ppt) during 2003–2007. Similar contrasting behaviour is also seen for several other models at lesser distinction, e.g. GEOS-Chem, NIES-08i lie below the fitted line for CH4, but lie above the fitted line for CH3CCl3. The CH4 IH gradients are best reproduced using the CH4INV emissions: deviations are within 5 ppb for 7 models (Fig. 10). Taking into account the IH gradient of all three species, TM5/CCAM, ACTM/IMPACT,1×1/PCTM and LMDZ/NIES-08 showed systematically higher, similar and lower IH gradients, respectively, compared to the observations.

Two models submitted simulations using different OH, which can be used to further investigate the sensitivity of the simulated CH4 and CH3CCl3 IH gradients to the IH difference in OH. The NH/SH ratio of hemispheric total tropospheric OH are (1) 1.32 for ACTM_OH and (2) 1.11 for GEOS-Chem_DOH, while that (Spivakovsky et al., 2000) used in ACTM is 0.99. The observed CH4 IH gradient is 100.97 ppb, while the simulated gradients from ACTM and ACTM_OH are 99.99 and 87.45 ppb, respectively. In order to simulate the observed CH4 IH gradient precisely, the NH/SH OH ratio for ACTM would need to be 0.97 [(= 0.99 + (1.32 − 0.99) × (99.99−100.97)) / (100.97−87.45)]. This is close to the NH/SH OH ratio derived by Spivakovsky et al. (2000). Note also that the ACTM simulations of both CH3CCl3 and SF6 agree very well with the observations. Similarly, to simulate the observed CH4 IH gradient, the NH/SH OH ratio for GEOS-Chem model should be 1.15 [(= 0.99 + (1.11 − 0.99) × (107.53−100.97)) / (100.97−107.53)]. Both GEOS-Chem model versions simulate greater IH gradients also for CH3CCl3, a species that has been used for benchmarking tropospheric OH concentrations and distributions. However, the GEOS-Chem sim-
suggests that the eight-years averaged growth rates at the surface sites for the different models are, as expected, inversely proportional to the calculated photochemical destruction. However, this relationship appears loose, particularly for 2000–2007, when the models approach steady state (correlation coefficient, \( r = -0.42 \) for all models, but increase to \(-0.82\) and \(-0.61\) for 1990s and 2000s, respectively, by excluding NIES-08i). Possible explanations are investigated in Fig. 11b, using the vertical gradients in the equatorial lower stratosphere (CH\(_4\) at 100 mb – CH\(_4\) at 10 mb; zonal average for 5°S–5°N latitudes). Models showing greater gradients have slower Brewer-Dobson circulation, and thus stronger trapping of CH\(_4\) in the troposphere, resulting in faster CH\(_4\) destruction and smaller growth rate because CH\(_4\) lifetime in the troposphere is an order of magnitude shorter than that in the stratosphere. Although these relationships are again quite loose during the 1990s, as the models attain their steady state, statistically significant correlations (\( r = -0.69 \)) are found for the period of 2000–2007 between the CH\(_4\) growth rates and vertical gradients in the lower stratosphere. While the growth rates decrease from the 1992–1999 period to the 2000–2007 period in all models, the modeled gradients remain remarkably constant. Based on this analysis we suggest that the simulated concentration growth rates at the surface sites are linked to the troposphere to stratosphere transport rate of CH\(_4\).

Figure 12 shows the temporal variability in the estimated lifetimes (Eq. 5) for CH\(_4\)_CTL and CH\(_3\)CCl\(_3\) (ref. Table 2 for time-averaged model specific lifetimes). The median CH\(_4\) lifetime due to atmospheric loss processes (Reactions R1–R3) is 9.99 ± 0.08 yr and CTL (9.99 ± 0.08 yr) fluxes agree within their interannual variability. A comparison of CH\(_3\)CCl\(_3\) lifetimes calculated using Eq. (5) and gridded photochemical destruction rates is shown in Fig. S22 (ACTM only).
Although the simulated zonal mean seasonal cycles of CH4 and CH3CCl3 lifetimes during the 1990s is an order of magnitude higher than that in the 2000s. We find up to 5% variability in the modeled CH3CCl3 lifetimes during 1992–1999, a period with substantial emissions. During 2000–2007, when the emissions of CH3CCl3 dropped significantly, less than 0.4% variability is simulated. This implies that inversions to estimate OH from CH3CCl3 observations are less certain since 2000, a finding in good agreement with Montzka et al. (2011).

4 Further work and data accessibility

For this analysis we used results of chemical tracer simulations at only 8 selected sites with measurements of atmospheric CH4, SF6 and CH3CCl3, which is a very small subset of the 280 surface sites for which output is available. In addition, vertical profiles of chemical tracers and several meteorological parameters have been archived at 115 sites. More analyses on the basis of the TransCom-CH4 simulations are planned focusing on (1) CH4 vertical profiles measured using aircraft, (2) analysis of vertical column averaged CH4 concentrations using TCCON observations, (3) using increase, decrease and exponential decay of CH3CCl3 for optimizing tropospheric OH abundance. We also welcome use of this data set by the measurement community. In an effort towards ease of access, time series at a subset of surface sites are archived at JAMSTEC (http://ebrcpa.jamstec.go.jp/dav/prabir/transcom-ch4/sites_data/). Information on how to access the full dataset is available in the experimental protocol, archived on this website (http://transcom.project.asu.edu/T4_methane.php).

In addition to the site-specific data, gridded output at monthly intervals at the model horizontal resolution at standard pressure levels are archived for the period 1990–2007. Afternoon averages (12:00–15:00 LT – Local Time) at daily intervals are also archived for the period 2001–2007. We believe these sets of model output, their extension to recent years, will be useful for comparing the model simulations with satellite observations (SCIAMACHY, AIRS, GOSAT) and aircraft observations (e.g. HIPPO – HIAPER Pole-to-Pole Observations of carbon cycle and greenhouse gases study, Wofsy et al., 2011).

5 Summary and conclusions

We analyzed concentration time series of CH4, CH3CCl3, SF6 and 222Rn simulated by 16 chemistry-transport models as part of the TransCom-CH4 intercomparison experiment. We focused the analysis on the model-to-model differences in:

1. The vertical redistribution of tracers, based on 222Rn simulations and comparisons to CH4 satellite observations in the upper troposphere and lower stratosphere.

2. Large-scale interhemispheric (IH) transport, by comparing modeled and observed IH gradients of SF6, CH4 and CH3CCl3.

3. Simulated seasonal cycles, by comparing to observed seasonal cycles at remote background stations.

4. Inter-annual variations in the simulated CH4 growth rate, by focusing on the results of six different CH4 emission time-lines.

5. The role of removal by OH on the simulated CH4 and CH3CCl3 concentrations.

The main conclusions can be summarized as follows:

i. Although the simulated zonal mean 222Rn concentrations agree between models, significant differences are observed in regions of deep cumulus convection, e.g. the south Asian summer monsoon domain. Unfortunately, observational evidence to check the model behavior is lacking. Models also differ in the simulated height of large troposphere-stratosphere concentration gradients of CH4. Compared to CH4 satellite observations in the upper troposphere, most models appear to be too diffusive around the tropical tropopause.

ii. The IH exchange time, calculated from the simulated SF6 distributions, ranges from 1.79 to 1.17 yr (average over 1996–2007) for the different models. The model average value of 1.39 yr is in close agreement with earlier studies and observational evidence. Models that show faster IH exchange for SF6 also exhibit faster exchange (smaller IH gradients) for CH4 and CH3CCl3. This multi-tracer evidence provides clear directions for the improvement of specific models. The IH exchange time calculated from the simulated and observed SF6 time series suggest an acceleration of the IH exchange in the 2000s compared to the 1990s. However, a sensitivity study using 3 different SF6 emission scenarios show that the acceleration is caused by an increased share of SF6 emissions in the tropical land regions, compared to the midlatitudes, of the NH as well as non-steady state IH gradients in SF6 concentrations. The estimated IH exchange time remained constant over the period of our analysis (1996–2007) for a time invariant SF6 emission scenario.

iii. All models reproduce the observed seasonal cycles of CH4 and CH3CCl3 at background sites very well. The
simulated CH$_4$ seasonal cycles depend on the destruction by reaction with OH, concealing the seasonality of the underlying emissions. The simulated seasonal cycles of CH$_4$ are influenced by the photochemical destruction by OH: without destruction, the simulated seasonal cycles would more strongly reflect the seasonal cycles present in the emission scenarios. Two of the six simulated CH$_4$ scenarios (CH$_4$INV and CH$_4$EXTRA) show a higher correlation with the observed seasonal cycle.

iv. A set of six CH$_4$ flux representations was used to investigate the role of specific processes in reproducing the observed interannual variations in the CH$_4$ growth rate. The control emission case (CH$_4$CTL) without a significant increase in anthropogenic emissions and no interannual variability in natural emissions for the period 1990–2007 reproduces the declining growth rate in the 1990s, followed by the stabilization in the 2000s. Inclusion of interannual variation in emissions from the wetland and forest fires (CH$_4$EXTRA) most successfully simulates the observed interannual variations in CH$_4$. It was also suggested that the higher tropospheric temperatures during the 1997/1998 El Niño resulted in larger CH$_4$ destruction, whereas the observations clearly show a rise in CH$_4$ concentration. To match the observations, either enhanced emissions are required (as in CH$_4$EXTRA), or less than average OH should have been present in this period.

v. The simulation of CH$_3$CCl$_3$ is used to check the consistency of the employed OH abundance and distribution. Two models used an alternative OH field next to the prescribed field. Generally, models that simulate a low abundance of CH$_3$CCl$_3$ also simulate a low abundance of CH$_4$. However, there are exceptions, which indicates that CH$_4$ loss due to OH in the troposphere is not the only cause of the modeled CH$_4$ differences. Thus, horizontal and vertical transport differences may also be important.

vi. Further analysis reveals that the simulated CH$_4$ growth rate shows (weak) correlations with the modeled vertical gradient in the equatorial lower stratosphere. This suggests that differences in vertical mixing of the emissions and in stratosphere-troposphere exchange are the main causes of the model-to-model differences. Next to the interhemispheric transport in models, this issue requires further analysis, e.g. based on the archived model output.

vii. Finally, the multi-model lifetime estimates for CH$_4$ and CH$_3$CCl$_3$ were found to be fairly constant over the simulation period with median values of 9.99 ± 0.08 and 4.61 ± 0.13 for the period 1992 to 2007. This underscores the fact that OH (assumed constant in the simulations) is the driving factor in the budgets of these gases, and that transport and temperature (affecting the reaction rate) differences play a smaller role. We find net CH$_4$ emissions to the atmosphere of 513 ± 9 and 514 ± 14 Tg CH$_4$ yr$^{-1}$ (soil sink subtracted) for the 1990s and 2000s, respectively, are consistent with the atmospheric losses accounted for due to OH, O$_1$D and Cl in order to simulate CH$_4$ concentrations and growth rates at the surface sites.

Supplementary material related to this article is available online at: http://www.atmos-chem-phys.net/11/12813/2011/acp-11-12813-2011-supplement.pdf.

Acknowledgements. We thank Ingeborg Levin for useful discussions on the possible causes for the decrease IH exchange time as derived from the measurements and control simulation in TransCom-CH4. This work is partly supported by JSPS/MEXT KAKENHI-A grant number 22241008. Generous support of the NIES team with the disk space and FTP services is greatly appreciated. Support of Lori Bruhwiler, Daniel Jacob, Philippe Ciais, and Frederic Chevallier is appreciated in the early stages of this experiment. We thank the ACE-FTS, AGAGE, GMD/ESRL and HALOE teams for observations; EDGAR, GISS/NASA, GFEDv2 (Guido van der Werf), IPSL_WETL (Bruno Ringeval), RIGC_Rice (Xiaoyuan Yan) for emission inventories. Without these datasets TransCom-CH4 experiment could not be completed. The ACE mission is funded primarily by the Canadian Space Agency. Annemarie Fraser is supported by the UK Natural Environment Research Council National Centre for Earth Observation. ACCESS uses the Met Office Unified Model™ (UK) with a CSIRO land-surface scheme (CABLE) and the help of Met Office and CSIRO staff, particularly Chris Jones and Fiona O’Connor, Martin Dix and Eva Kowalczyk, was much appreciated. We acknowledge the work of John McGregor and Marcus Thatcher in the development of CCAM. ACCESS and CCAM simulations were undertaken as part of the Australian Climate Change Science Program and used the NCI National Facility in Canberra, Australia. The contribution by the LLNL authors was prepared under Contract DE-AC52-07NA27344, with different parts supported by the IMPACTS project funded by the US DOE (BER) and project (07-ERD-064) funded by the LDRD program at LLNL. Ronald Prinn and Matthew Rigby are supported by NASA-AGAGE Grant NNX11AF17G to MIT. The TOMCAT work at University of Leeds was supported by NERC/NCEO. We sincerely thank two anonymous reviewers and James Wang for providing us constructive comments and helpful suggestions.

Edited by: B. N. Duncan

References


Gent, P. R., Yeager, S. G., Neale, R. B., Levis, S., and Bailey, D. A.: Improvements in a half degree atmosphere/land version of the...


P. K. Patra et al.: TransCom model simulations of CH\textsubscript{4} and related species


Yan, X., Akiyama, H., Yagi, K., and Akimoto, H.: Global estima-