

Reconciling the changes in atmospheric methane sources and sinks between the Last Glacial Maximum and the pre-industrial era

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Received 2 September 2011; revised 18 October 2011; accepted 24 October 2011; published 8 December 2011.

[1] We know from the ice record that the concentration of atmospheric methane, [CH₄], at the Last Glacial Maximum (LGM) was roughly half that in the pre-industrial era (PI), but how much of the difference was source-driven, and how much was sink-driven, remains uncertain. Recent developments include: a higher estimate of the LGM-PI change in methane emissions from wetlands—the dominant, natural methane source; and the possible recycling of OH consumed in isoprene oxidation—the principal methane sink. Here, in view of these developments, we use an atmospheric chemistry-transport model to re-examine the main factors affecting OH during this period: changes in air temperature and emissions of non-methane volatile organic compounds from vegetation. We find that their net effect was negligible (with and without an OH recycling mechanism), implying the change in [CH₄] was almost entirely source driven—a conclusion that, though subject to significant uncertainties, can be reconciled with recent methane source estimates.

Citation: Levine, J. G., E. W. Wolff, A. E. Jones, L. C. Sime, P. J. Valdes, A. T. Archibald, G. D. Carver, N. J. Warwick, and J. A. Pyle (2011), Reconciling the changes in atmospheric methane sources and sinks between the Last Glacial Maximum and the pre-industrial era, *Geophys. Res. Lett.*, 38, L23804, doi:10.1029/2011GL049545.

1. Introduction

[2] Methane (CH₄) is an important atmospheric constituent on account of its potency as a greenhouse gas and its influence on the tropospheric oxidizing capacity. We know from the ice record that its concentration, [CH₄], at the Last Glacial Maximum (LGM; 21kyr before present) was roughly half that in the pre-industrial era (PI; ~200 yr before present): about 360 ppbv compared to around 700 ppbv [e.g., *Loulergue et al.*, 2008]. But how much of this difference was source-driven and how much was sink-driven remains uncertain. Bottom-up model studies (employing dynamic global vegetation models) suggest the reduction in natural CH₄ sources at the LGM, the dominant one being wetlands, can at most account for roughly half the reduction in [CH₄]

[e.g., *Valdes et al.*, 2005; *Kaplan et al.*, 2006]. These studies therefore suggest an increase in CH₄ sinks, of which oxidation by the hydroxyl radical (OH) is by far the greatest, is needed to explain the remainder of the reduction, appealing to an increase in [OH] due to a reduction in non-methane volatile organic compound (NMVOC) emissions. Recent calculations by *Weber et al.* [2010], however, suggest the reduction in CH₄ emissions from wetlands at the LGM was greater than previously simulated, implying a greater role for CH₄ sources. Meanwhile, the proposed recycling of OH consumed in the oxidation of isoprene [*Lelieveld et al.*, 2008]—the most abundant biogenic NMVOC [e.g., *Guenther et al.*, 2006]—could alter our understanding of the influence a reduction in NMVOC emissions at the LGM had on OH, and hence the size of the sink-driven component. Here, we use an atmospheric chemistry-transport model to (1) re-examine the main factors affecting the oxidizing capacity between the LGM and the PI, (2) estimate their net effect on [CH₄], subject to an established, conventional isoprene oxidation scheme, with and without the addition of a possible OH recycling mechanism, and (3) show how these estimates can be reconciled with both the findings of *Weber et al.* [2010] and the observed change in [CH₄].

[3] *Chappellaz et al.* [1993] estimated, based on a reconstruction of vegetation, that CH₄ emissions from wetlands were 44% lower at the LGM than in the PI, underpinning a 33% reduction in total CH₄ emissions (including emissions from termites, biomass burning and the ocean). ‘Top-down’ Monte-Carlo calculations, designed to explain ice-core measurements of both [CH₄] and δ¹³CH₄, suggest a reduction in wetland emissions of between 41 and 57% is plausible [see *Fischer et al.*, 2008, Table S2]. But until recently, bottom-up model studies have put the figure significantly lower, between effectively no change [*Kaplan et al.*, 2006] and a reduction of 27% [*Valdes et al.*, 2005], yielding reductions in total CH₄ emissions of just 16% and 23% respectively. *Weber et al.* [2010], however, have calculated reductions in wetland emissions of between 35 and 42%. Had *Valdes et al.* [2005] and *Kaplan et al.* [2006] simulated a 35–42% reduction in wetland emissions, they would have calculated reductions in total CH₄ emissions of 30–35% and 38–43% respectively. This opens up new possibilities regarding the explanation of the reduction in [CH₄] at the LGM. Allowing for the small positive feedback between the concentration and atmospheric lifetime of CH₄ [see, e.g., *Prather*, 2007], the largest of these reductions in total CH₄ emissions, one of 43%, could account for practically the entire 49% reduction in [CH₄], from 700 to 360 ppbv, without any increase in oxidizing capacity (besides that which would have been induced by the reduction in [CH₄]). We note that a recent

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Table 1. Main Features of Experiments with p-TOMCAT^a

Experiment	T _{Humidity}	T _{Kinetics}	E _{NMVOCS}	OH Recycling	Resolution	[CH ₄] _{AntBL} (ppbv)	τ _{CH₄} (years)
A	PI	PI	PI	None	2.8° × 2.8°; 31 lev	714	8.5
B	LGM	PI	PI	None	2.8° × 2.8°; 31 lev	763 (+7%)	9.1 (+7%)
C	LGM	LGM	PI	None	2.8° × 2.8°; 31 lev	840 (+18%)	10.0 (+18%)
D	PI	PI	None	None	2.8° × 2.8°; 31 lev	558 (-22%)	6.7 (-22%)
E	LGM	LGM	None	None	2.8° × 2.8°; 31 lev	637 (-11%)	7.6 (-11%)
F	LGM	LGM	LGM	None	2.8° × 2.8°; 31 lev	692 (-3%)	8.3 (-3%)
G	PI	PI	PI	None	5.6° × 5.6°; 19 lev	715	8.6
H	LGM	LGM	LGM	None	5.6° × 5.6°; 19 lev	703 (-2%)	8.5 (-2%)
I	PI	PI	PI	Fixed	5.6° × 5.6°; 19 lev	703	8.5
J	LGM	LGM	LGM	Fixed	5.6° × 5.6°; 19 lev	689 (-2%)	8.3 (-2%)
K	PI	PI	PI	T-dep.	5.6° × 5.6°; 19 lev	707	8.5
L	LGM	LGM	LGM	T-dep.	5.6° × 5.6°; 19 lev	692 (-2%)	8.3 (-2%)

^aT_{Humidity} and T_{Kinetics} refer to the temperatures influencing the humidity and chemical kinetics; E_{NMVOCS} refers to the NMVOC emissions from vegetation (as specified in Table 2); and ‘fixed’ and ‘T-dep.’ refer to fixed and temperature-dependent rate coefficients in the OH-recycling reactions (1 and 2). The [CH₄]_{AntBL} and τ_{CH₄} calculated in each experiment are given (see Section 2 for definitions), as are the percentage changes in these quantities in each sensitivity experiment relative to the corresponding PI model run; see text for details.

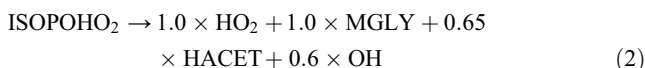
study by *Singarayer et al.* [2011] places a smaller emphasis on the role of CH₄ sinks than the studies of *Valdes et al.* [2005] and *Kaplan et al.* [2006].

[4] In principle, many factors could have affected the oxidizing capacity (see discussion in Section 4), but only two are identified in the literature as having had a significant influence between the LGM and the PI: the air temperature and NMVOC emissions from vegetation. The lower air temperatures at the LGM would have been accompanied by lower humidities, and hence lower [OH], as OH is formed from the reaction between excited oxygen atoms (O(¹D)) and water vapor. *Martinerie et al.* [1995] estimated that, all else being equal, this would have increased the lifetime of CH₄, τ_{CH₄}, by 7%. The lower air temperatures would have also been responsible for slower rates of reaction between OH and CH₄, which *Valdes et al.* [2005] estimated, together with the lower [OH], would have led to a 13% increase in τ_{CH₄}; *Harder et al.* [2007] illustrated the sensitivity that τ_{CH₄} shows to the degree of cooling at the LGM, calculating increases of 8%, 24% and 57% subject to three different representations of the glacial climate, based on data from *Climate: Long-Range Investigation, Mapping and Prediction (CLIMAP)* [1981], (see *Harder et al.* [2007] for details) and *Webb et al.* [1997] respectively. The physical effects of the lower temperatures at the LGM would thus have been to increase, not decrease, [CH₄]. However, the cooler, drier climate of the LGM would have also seen lower NMVOC emissions from vegetation [e.g., *Adams et al.*, 2001; *Lathière et al.*, 2005; *Valdes et al.*, 2005; *Kaplan et al.*, 2006]. NMVOCS, such as isoprene, compete with CH₄ for reaction with OH, and hence a reduction in their emissions at the LGM would have tended to increase [OH] and reduce τ_{CH₄}—by 17% according to *Valdes et al.* [2005]. There is, however, still considerable uncertainty regarding the degree to which the oxidation of isoprene ‘consumes’ OH; large discrepancies have been observed between modeled and measured [OH] in regions of high isoprene (and low nitrogen-oxide; NO_x = NO + NO₂) emissions [*Lelieveld et al.*, 2008; *Hewitt et al.*, 2010], and various so-called OH-recycling mechanisms have been proposed to resolve these [see, e.g., *Archibald et al.*, 2010]. One such mechanism, based on the calculations of *Peeters et al.* [2009], shows significant potential [*Archibald et al.*, 2010; *Stavrakou et al.*, 2010], and we therefore test

the sensitivity of our results, obtained with the conventional Mainz Isoprene Mechanism [*Pöschl et al.*, 2000], to its inclusion.

2. Model Experiments

[5] We carry out twelve experiments with the Cambridge parallelised-Tropospheric Offline Model of Chemistry and Transport (p-TOMCAT): the main features of these are summarized in Table 1. p-TOMCAT is a 3D global Eulerian model, run at a resolution of 2.8° × 2.8° on 31 levels (unless otherwise stated), and driven by wind, temperature and humidity data from the operational analyses of the European Centre for Medium-range Weather Forecasts (ECMWF). The model’s chemistry includes the HO_x/NO_x chemistries of methane, ethane, propane and isoprene, with the oxidation of isoprene following the Mainz Isoprene Mechanism [*Pöschl et al.*, 2000]; see *Levine et al.* [2011] for more details on both the physical and the chemical aspects of the model. In a number of the experiments, however, we add two reactions (equations (1) and (2)); ISOPOHO₂ = isoprene hydroxy peroxy radical, MACR = lumped C4 carbonyl, MGLY = methyl glyoxal, and HACET = hydroxyl-acetone), which together comprise an OH recycling mechanism based on the calculations of *Peeters et al.* [2009], previously implemented in the UKCA global chemistry-climate model [*Archibald et al.*, 2011]. To assess the sensitivity of our results to the temperature dependence of these additional reactions, we employ both fixed rate coefficients, as implemented by *Archibald et al.* [2011] ($k_1 = 4.0 \times 10^{-3} \text{ s}^{-1}$ and $k_2 = 8.0 \times 10^{-2} \text{ s}^{-1}$), and temperature-dependent ones, based on the calculations of *Peeters et al.* [2009] ($k_1 = 2.1 \times 10^{11} e^{-8993/T}$ and $k_2 = 5.9 \times 10^9 e^{-7366/T}$, where T is the temperature in K).



[6] Experiment A (referred to by *Levine et al.* [2011] as the ‘PI model run’) employs PI trace gas emissions [*Valdes et al.*,

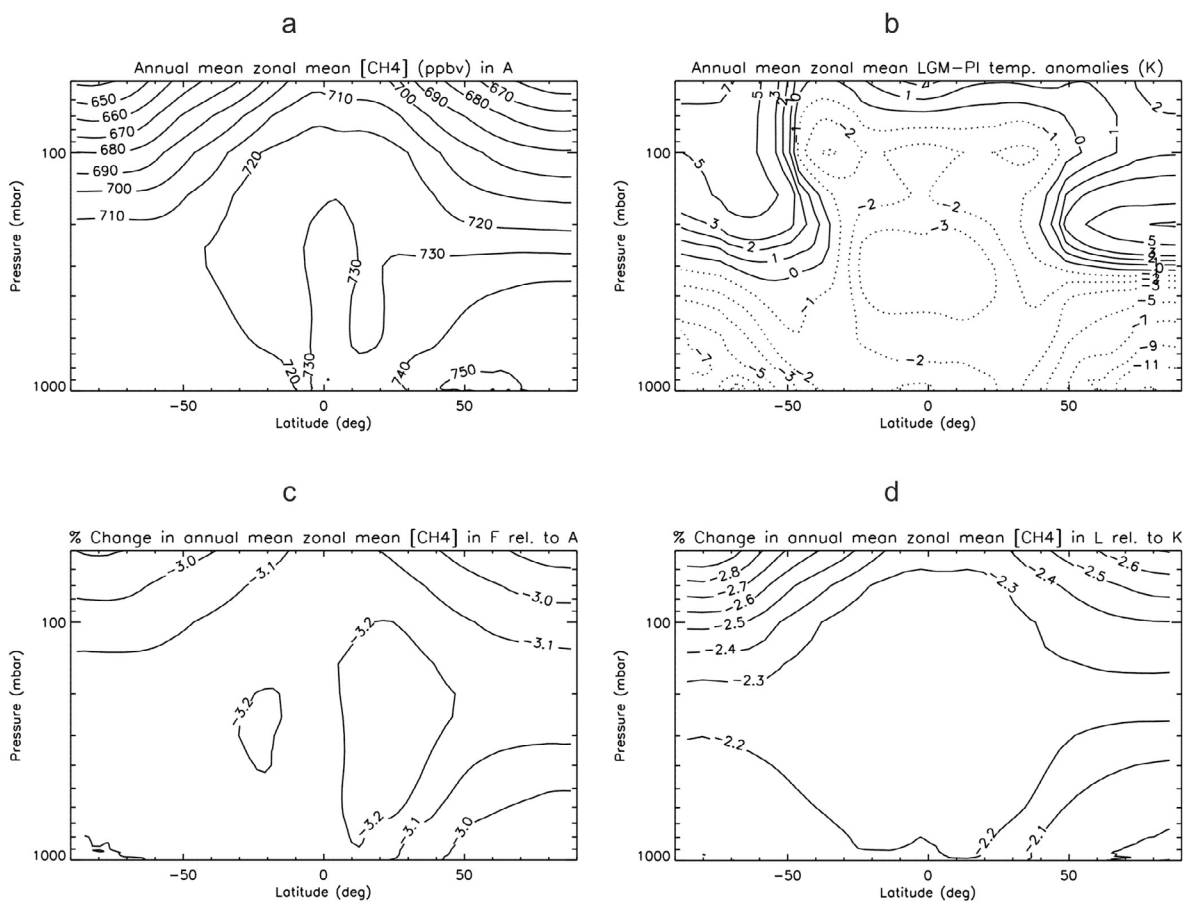


Figure 1. (a) Annual mean zonal mean [CH₄] in experiment A, (b) annual mean zonal mean LGM-PI temperature anomalies, and (c, d) percentage changes in annual mean zonal mean [CH₄] in experiments F and L, relative to A and K.

2005] and present-day meteorology (assuming recent meteorological analyses adequately represent the meteorology of the PI). The emissions include PI emissions of methane, which remain unchanged throughout the experiments in order to explore the impacts of changes in air temperature and NMVOC emissions (alone) on the oxidizing capacity. In experiment B, we modify the humidity in the model in line with LGM air temperatures, recalculating the saturated volume mixing ratio of water vapor with respect to ice according to *Marti and Mauersberger [1993]* and adjusting the water vapor volume mixing ratio so as to conserve relative humidity. In experiment C, we also use LGM air temperatures to influence the chemical kinetics. The LGM air temperatures are obtained by combining LGM-PI temperature

anomalies calculated with the Hadley Centre Atmosphere-only general circulation model, HadAM3 (see auxiliary material), and the ECMWF air temperatures employed in experiment A; the zonal-mean temperature anomalies are illustrated in Figure 1b.¹

[7] In experiment D (referred to by *Levine et al. [2011]* as ‘Sink 2’), we remove all emissions from vegetation (as specified in Table 2), and in experiment E, we do so subject to LGM air temperatures (affecting both the humidity in the model and the chemical kinetics). In experiment F, instead of removing all emissions from vegetation, we reduce them

¹Auxiliary materials are available in the HTML. doi:10.1029/2011GL049545.

Table 2. Trace-Gas Emissions^a

Trace Gas	Biomass Burning	Oceans	Vegetation	Soils	Lightning	Wetlands	Termites	Total
NO ₂	1.4	-	-	5.1	4.8	-	-	11.3
CH ₄	11.0	13.0	-	-	-	147.9	27.0	198.9
CO	100.0	50.0	150.0 (100.5)	-	-	-	-	300.0 (250.5)
C ₂ H ₆	0.7	-	3.5 (2.3)	-	-	-	-	4.2 (3.0)
C ₃ H ₈	0.2	0.5	3.5 (2.3)	-	-	-	-	4.2 (3.0)
CH ₃ COCH ₃	0.1	-	20.0 (13.4)	-	-	-	-	20.1 (13.5)
C ₅ H ₈	-	-	673.7 (258.9)	-	-	-	-	673.7 (258.9)
HCHO	0.3	-	-	-	-	-	-	0.3
CH ₃ CHO	0.8	-	-	-	-	-	-	0.8

^aTrace-gas emissions are expressed in Tg molecular mass year⁻¹, with the exception of NO₂ emissions, which are expressed in Tg N year⁻¹. Numbers in parentheses refer to LGM emissions; all others refer to PI emissions [see *Valdes et al., 2005*].

in line with LGM estimates [Valdes *et al.*, 2005], based on calculations with the University of Sheffield Dynamic Global Vegetation Model [Woodward *et al.*, 1995; Beerling and Woodward, 2001] (see Table 2). Experiments I (K) and J (L) are repeats of experiments A and F, subject to OH recycling with fixed (temperature-dependent) rate coefficients, k_1 and k_2 . They are carried out at a lower resolution of $5.6^\circ \times 5.6^\circ$ on 19 levels, having first repeated experiments A and F at this resolution (experiments G and H) to verify that the annual-mean [CH₄] in the Antarctic boundary layer, [CH₄]_{AntBL} (defined to comprise all boxes in the lowest level of the model south of 70°S), and τ_{CH_4} (defined as the annual-mean tropospheric burden of CH₄ divided by the annual CH₄ emission rate) are insensitive to the resolution.

3. Results

[8] The [CH₄]_{AntBL} and τ_{CH_4} calculated in each experiment (once the model has reached equilibrium; see Levine *et al.* [2011] for definition) are given in Table 1. In experiments B–F, [CH₄]_{AntBL} and τ_{CH_4} show very similar percentage changes relative to their values in experiment A; similarly, [CH₄]_{AntBL} and τ_{CH_4} show very similar percentage changes in experiments H, J and L, relative to their values in experiments G, I and K, respectively. We therefore refer almost exclusively to changes in τ_{CH_4} . Like Martinerie *et al.* [1995], we find that the lower humidity of the LGM leads to a 7% increase in τ_{CH_4} (experiment B). Combined with the changes in chemical kinetics, we calculate an 18% increase in τ_{CH_4} (experiment C); Valdes *et al.* [2005] calculated a 13% increase in τ_{CH_4} in response to the cooling at the LGM, with trace-gas emissions fixed at LGM levels.

[9] In contrast, the removal of all NMVOC emissions from vegetation leads to a 22% reduction in τ_{CH_4} (experiment D); this is an estimate of the maximum influence a reduction in NMVOC emissions at the LGM could have had on τ_{CH_4} . Combined with the effects of the changes in air temperature, we calculate an 11% reduction in τ_{CH_4} (experiment E), and when instead of removing all NMVOC emissions from vegetation, we reduce them in line with LGM estimates, we calculate a reduction of just 3% (experiment F). The percentage difference in annual-mean zonal-mean [CH₄] in experiment F, relative to experiment A, is illustrated in Figure 1c.

[10] Repeating experiments A and F at lower resolution—experiments G and H—we calculate similar values of τ_{CH_4} (8.6 years in G c.f. 8.5 years in A, and 8.5 years in H c.f. 8.3 years in F) and similar values of [CH₄]_{AntBL} (715 ppbv in G c.f. 714 ppbv in A and 703 ppbv in H c.f. 692 ppbv in F). We therefore also calculate a similar percentage change in the two quantities, τ_{CH_4} and [CH₄]_{AntBL}, in response to the changes in air temperature and NMVOC emissions at the LGM (−2% in H relative to G c.f. −3% in F relative to A). Our results appear to be reasonably insensitive to model resolution.

[11] At the lower resolution, the addition of OH recycling to our PI climate/emission scenario (with both fixed and temperature-dependent rate coefficients) reduces τ_{CH_4} by between 1 and 2%, from 8.6 years in experiment A to around 8.5 years in experiments I and K. This is similar to the 3% reduction calculated by Cooke [2010] with STOCHEM—

the UK Meteorological Office Global Three-Dimensional Lagrangian model—but contrasts with the 35% reduction calculated by Archibald *et al.* [2011] with UKCA. A further study is needed to understand the differences between the models in this regard. Here, we are primarily interested in the influence that recycling has on the change in τ_{CH_4} in response to the changes in air temperature and NMVOC emissions at the LGM, relative to the PI. Subject to recycling (with both fixed and temperature-dependent rate coefficients), we find that the changes at the LGM result in just a 2% reduction in τ_{CH_4} (experiments J and L relative to experiments I and K respectively). The influence of recycling in our calculations thus appears to be small, and its temperature-dependence non-critical. Figure 1d illustrates the percentage difference in annual-mean zonal-mean [CH₄] in experiment L, relative to experiment K.

4. Discussion

[12] Our experiments with the Cambridge p-TOMCAT model suggest the net effect on τ_{CH_4} of the changes in air temperature and NMVOC emissions at the LGM, relative to the PI, was negligible: their separate effects, though substantial, were roughly equal and opposite, resulting in a net change in τ_{CH_4} of between 2 and 3%. This finding holds irrespective of whether we include an OH recycling mechanism based on the calculations of Peeters *et al.* [2009], previously shown to reduce modeled/measured [OH] discrepancies [Archibald *et al.*, 2010; Stavrakou *et al.*, 2010] of the sort identified by Lelieveld *et al.* [2008] and Hewitt *et al.* [2010]. If the changes in air temperature and NMVOC emissions were the only factors to significantly influence the oxidizing capacity during this period, the implication is that the LGM-PI change in [CH₄] was almost entirely source-driven—a conclusion that, as outlined in Section 1, is consistent with recent estimates of CH₄ source strengths if CH₄ emissions from wetlands changed in line with Weber *et al.*'s [2010] highest estimate.

[13] In principle, other factors could have affected the oxidizing capacity. These include changes in: albedo, cloudiness, stratospheric ozone and NO_x emissions. The increase in the amount of land and sea covered by ice at the LGM would have been accompanied by an increase in albedo at high latitudes, and hence an increase in the amount of UV radiation available here to photolyze tropospheric ozone to produce the excited oxygen atom, O(¹D)—a key step in the production of OH. This would have tended to increase [OH] and reduce τ_{CH_4} and [CH₄]. However, given that only around 25% of CH₄ is oxidized outside of the tropics, and only a small fraction of this at high latitudes [see, e.g., Lawrence *et al.*, 2001; Labrador *et al.*, 2004], it is unlikely to have had a significant effect on τ_{CH_4} globally. Similarly, changes in cloudiness, which would also affect the amount of UV radiation available to photolyze tropospheric ozone, are unlikely to have had a significant influence; Ramstein *et al.*, 1998, Figure 13] estimated that zonal-mean cloud cover changed by 5–10% at high latitudes and less than or similar to 1% at all other latitudes. Likewise regarding stratospheric ozone: a series of model studies has concluded that the total ozone column changed, at all latitudes, by less than or similar to 2% [see Pinto and Khalil, 1991; Crutzen and Brühl, 1993; Martinerie *et al.*, 1995; Karol *et al.*, 1995; Rind

et al., 2009], though we note that these studies were only loosely constrained by ice-core measurements of [CH₄], [CO₂] and [N₂O].

[14] Though NO_x emissions play an important part in determining the oxidizing capacity, governing the formation of tropospheric ozone accompanying the oxidation of CH₄ and NMVOCs, it remains uncertain how they changed between the LGM and the PI. It is estimated that soils, lightning and biomass burning accounted for roughly 45%, 40% and 15%, respectively, of total NO_x emissions in the PI [see, e.g., Valdes et al., 2005] (Table 2). Valdes et al. [2005] estimated that NO_x emissions from soils were 12% greater at the LGM whilst those from lightning were 19% lower, resulting in a small reduction in total NO_x emissions that would have led, had all else been equal, to a 6% increase in τ_{CH_4} at the LGM. They did so, however, using only an empirical (c.f. process-based) model of soil NO_x fluxes [Yienger and Levy, 1995], and assuming NO_x emissions from lightning, to which τ_{CH_4} is especially sensitive (increasing/decreasing by more than a year in response to a 2.5Tg N year⁻¹ decrease/increase in emissions from this source [Wild, 2007]), scaled with the amount of convective precipitation simulated with an atmospheric general circulation model. Their estimates are therefore subject to significant uncertainties. It is also unclear how NO_x emissions from biomass burning differed at the LGM: top-down Monte-Carlo calculations by Fischer et al. [2008] suggest there was little or no change in biomass burning whilst a recent global synthesis of charcoal records identifies the last glacial period (16–21kyr before present) as the period of least biomass burning in the last 21kyr [Power et al., 2008]. What we can say, however, is that any reduction in total NO_x emissions at the LGM would have tended to increase τ_{CH_4} , and thus increase the reduction in CH₄ emissions needed at the LGM to explain the reduction in [CH₄] observed.

[15] The changes in air temperature and NMVOC emissions that we have explored, not to mention the OH recycling mechanism employed [Archibald et al., 2011], are also subject to uncertainties. We have explored changes in air temperature amounting to a cooling at the LGM of about two degrees in the tropics (see Figure 1b). This lies roughly in the middle of the coolings exhibited by state-of-the-art climate models [see Braconnot et al., 2007, Figure 8], but as the latter exhibit a range of about a degree and are under-constrained by paleodata [see, e.g., Braconnot et al., 2007; Otto-Bliesner et al., 2009], it could be subject to an uncertainty of around $\pm 25\%$. The calculations of Harder et al. [2007] illustrate the sensitivity that τ_{CH_4} shows to the degree of cooling at the LGM. Though we would generally expect a greater (lesser) cooling to result in a greater (lesser) reduction in NMVOC emissions, and hence the continued offsetting of one effect on the oxidizing capacity against the other, an extreme cooling (well beyond the bounds of a $\pm 25\%$ uncertainty) could result in a net increase in τ_{CH_4} at the LGM. For instance, Harder et al. [2007] found that the cooling estimated by Webb et al. [1997] led to a 57% increase in τ_{CH_4} , whilst our calculations suggest the removal of all NMVOC emissions from vegetation would only reduce τ_{CH_4} by 22%. The uncertainty surrounding the changes in NMVOC emissions is still harder to assess. There are uncertainties inherent to:

the dynamic global vegetation model with which the emissions are simulated; the simulations of the atmospheric circulation with which this is forced; and the estimates of present-day emissions [see, e.g., Arneeth et al., 2008], on which estimates of past emissions implicitly depend. A further notable uncertainty surrounds the degree to which the ‘CO₂-suppression’ of isoprene emissions was reduced at the LGM relative to the PI [e.g., Arneeth et al., 2007; Wilkinson et al., 2009], as a result of the lower [CO₂] at the LGM [see, e.g., Lüthi et al., 2008]. Any reduction in CO₂-suppression at the LGM would have tended to increase τ_{CH_4} , and thus increase the reduction in CH₄ emissions needed at the LGM to explain the reduction in [CH₄] observed.

[16] In light of the significant uncertainties discussed above, we cannot assert that the LGM-PI change in [CH₄] was definitely entirely source-driven. However, this is the implication of our calculations and such a conclusion is consistent with recent estimates of CH₄ source strengths, if CH₄ emissions from wetlands changed in line with Weber et al.’s [2010] highest estimate—close to the top-down estimate of Fischer et al. [2008], based on Monte-Carlo calculations designed to explain ice-core measurements of both [CH₄], $\delta^{13}CH_4$, and $\delta D(CH_4)$; Section 1. Furthermore, any reduction at the LGM in the total amount of NO_x emitted, and/or the degree to which isoprene emissions were suppressed by CO₂, would only imply a still larger total CH₄-source change. We therefore conclude that it is plausible the LGM-PI change in [CH₄] was entirely source-driven, and that the changes in CH₄ sources and sinks between the LGM and the PI could be reconciled thus.

[17] **Acknowledgements.** This study is part of the British Antarctic Survey Polar Science for Planet Earth Programme and a contribution to the Dynamics of the Earth System and the Ice-core Record project, the latter being jointly funded by the Natural Environment Research Council’s programme, Quantifying and Understanding the Earth System, and the French Institut National des Sciences de l’Univers; we gratefully acknowledge their support. A.T.A., G.D.C., N.J.W. and J.A.P. acknowledge the support of the National Centre for Atmospheric Science. Finally, we express our thanks to two anonymous reviewers.

[18] The Editor thanks two anonymous reviewers for their assistance in evaluating this paper.

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