

**Methanotrophy in
Lake Kivu**

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Methanotrophy within the water column of a large meromictic tropical lake (Lake Kivu, East Africa)

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Abstract

The permanently stratified Lake Kivu is one of the largest freshwater reservoirs of dissolved methane (CH_4) on Earth. Yet CH_4 emissions from its surface to the atmosphere has been estimated to be 2 orders of magnitude lower than the CH_4 upward flux to the mixed layer, showing that microbial CH_4 oxidation is an important process within the water column. A combination of natural abundance carbon stable isotope analysis ($\delta^{13}\text{C}$) of several inorganic and organic carbon pools and $^{13}\text{CH}_4$ -labelling experiments was carried out during rainy and dry season to quantify (i) the contribution of CH_4 -derived carbon to the biomass, (ii) methanotrophic bacterial production (MBP), and (iii) methanotrophic bacterial growth efficiency (MBGE), defined as the ratio between MBP and gross CH_4 oxidation. We also investigated the distribution and the $\delta^{13}\text{C}$ of specific phospholipid fatty acids (PLFA), used as biomarkers for aerobic methanotrophs. Data revealed that methanotrophic organisms oxidized within the water column most of the upward flux of CH_4 to the mixed layer and a significant amount of CH_4 -derived carbon was incorporated into the microbial biomass in the oxycline. Maximal MBP rates were measured in the oxycline, suggesting that CH_4 oxidation was mainly driven by oxic processes. The MBGE was variable (2–50%) and negatively related to $\text{CH}_4:\text{O}_2$ molar ratios. Thus, a comparatively smaller fraction of CH_4 -derived carbon was incorporated into the cellular biomass in deeper waters, at the bottom of the oxycline where oxygen was scarce. The aerobic methanotrophic community was clearly dominated by type I methanotrophs and no evidence was found for an active involvement of type II methanotrophs in CH_4 oxidation in Lake Kivu. Vertically integrated over the water column, the MBP was equivalent to 16–58% of the average phytoplankton primary production. This relatively high magnitude of MBP, and the substantial contribution of CH_4 -derived carbon to the overall biomass in the oxycline, suggest that methanotrophic bacteria could potentially sustain a significant fraction of the pelagic food-web in the deep oligotrophic Lake Kivu.

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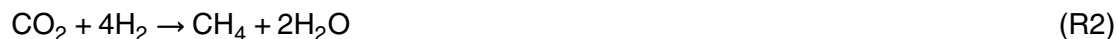
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1 Introduction

Although the atmospheric methane (CH₄) concentration is low compared to carbon dioxide (CO₂), CH₄ contributes significantly to the anthropogenic radiative forcing (18%) because of its 25 times higher global warming potential than CO₂ (Forster et al., 2007). CH₄ has several natural and anthropogenic sources and sinks, whereby natural and artificial wetlands are recognized as major CH₄ sources to the atmosphere (e.g. Kirschke et al., 2012). Bastviken et al. (2011) estimated that CH₄ emissions to the atmosphere from freshwater ecosystems (0.65 Pg C yr⁻¹ as CO₂ equivalent) would correspond to 25% of the global land carbon (C) sink (2.6 ± 1.7 Pg C yr⁻¹, Denman et al., 2007). Tropical regions are responsible for approximately half of the estimated CH₄ emissions from freshwater ecosystems to the atmosphere, although they have been consistently undersampled (Bastviken et al., 2011). Thus, more information on both the magnitude and controlling factors of CH₄ emissions from tropical inland waters are warranted. CH₄ is produced mainly in anoxic sediments by methanogenic archaea following two different pathways: acetoclastic methanogenesis (Reaction 1), using acetate produced from organic matter degradation, or CO₂ reduction (Reaction 2).



Although both methanogenic pathways may co-occur, CO₂ reduction is dominant in marine sediments, while acetate fermentation is the major pathway in freshwater sediments (Whiticar et al., 1986). CH₄ production rates are typically higher than CH₄ emission fluxes to the atmosphere, since aerobic and anaerobic microbial CH₄ oxidation within lacustrine sediments or in water columns are effective processes that limit the amount of CH₄ reaching the atmosphere, in particular when vertical CH₄ transport occurs mainly through diffusive transport, rather than through ebullition. A wide variety of electron acceptors can be used during microbial CH₄ oxidation, including but not limited to oxygen (O₂, Rudd et al., 1974) and sulphate (SO₄²⁻, Boetius

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et al., 2000). Micro-organisms performing aerobic CH₄ oxidation (Reaction 3) belong to the Proteobacteria phylum and anaerobic CH₄ oxidation coupled with SO₄²⁻ reduction (Reaction 4) is carried out by a syntrophic consortium of CH₄-oxidizing archaea and sulphate-reducing bacteria (Boetius et al., 2000), although up to now this process has not been frequently reported in freshwater ecosystems.



Methanotrophic organisms not only use CH₄ as electron donors but they are also able to incorporate a fraction of the CH₄-derived C into their biomass, and could therefore contribute to fuel the pelagic food web (Bastviken et al., 2003; Jones and Grey, 2011; Sanseverino et al., 2012). A recent study carried out in small boreal lakes (surface area < 0.01 km²) demonstrated that methanotrophic bacterial production (MBP) contributed to 13–52 % of the autochthonous primary production in the water column (Kankaala et al., 2013). But in spite of the potential importance of this alternative C source in aquatic ecosystems, direct measurements of MBP in lakes are still scarce. Also, the methanotrophic bacterial growth efficiency (MBGE), defined as the amount of biomass synthesized from CH₄ per unit of CH₄ oxidized, was found to vary widely in aquatic environments (15–80 % according to King, 1992; 6–77 % according to Bastviken et al., 2003), and moreover, little is known about the factors driving its variability.

Lake Kivu, located in a volcanic area, is one of the largest freshwater CH₄ reservoirs, with approximately 60 km³ (at standard temperature and pressure) dissolved in its permanently stratified water (Schmid et al., 2005). One third of the CH₄ accumulated in its deep waters is estimated to be produced via the acetoclastic pathway and two thirds by reduction of geogenic CO₂ (Schoell et al., 1988). Based on a modelling approach, Schmid et al. (2005) estimated that CH₄ production recently increased by threefold since the 1970s for a still unknown reason. However, the emission of CH₄ from surface waters to the atmosphere (0.038 mmol m⁻² d⁻¹, Borges et al., 2011) is several orders of magnitude lower than the upward flux of CH₄ to the mixed layer (9.38 mmol m⁻² d⁻¹,

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Pasche et al., 2009), showing that CH₄ oxidation prevents most of CH₄ to reach the lake surface waters.

Due to its inherent characteristics, the meromictic Lake Kivu offers an ideal natural laboratory to investigate the role of methanotrophy in large tropical lakes. In this study, we used the difference in C stable isotope abundance ($\delta^{13}\text{C}$) of different C sources to estimate the fraction of CH₄ inputs to the mixed layer from deep waters that is microbially oxidized within the water column, and to quantify the relative contribution of CH₄-derived C to the particulate biomass. Additionally, phospholipid fatty acids (PLFA) and their $\delta^{13}\text{C}$ signatures were analyzed to characterize the populations of methanotrophic bacteria present in the water column. We also carried out ¹³CH₄-labelling experiments to trace the incorporation of CH₄-derived C into the biomass (to quantify methanotrophic bacterial production) and its conversion to CO₂ (to quantify methanotrophic bacterial growth efficiency). Finally, stable isotope probing (SIP) of specific PLFA (SIP-PLFA) after ¹³C-CH₄ labelling allowed to characterize the bacterial populations active in methanotrophy.

2 Material and methods

2.1 Study site description and sampling

Lake Kivu (East Africa) is a large (2370 km²) and deep (maximum depth of 485 m) meromictic lake. Its vertical structure consists of an oxic and nutrient-poor mixed layer (seasonally variable depth, up to 70 m), and a permanently anoxic monimolimnion rich in dissolved gases (CH₄, CO₂) and inorganic nutrients (Damas, 1937; Degens et al., 1973; Schmid et al., 2005). Seasonal variations of the vertical position of the oxic–anoxic transition are driven by contrasting hygrometry and long wave radiation between rainy (October–May) and dry (June–September) seasons (Thiery et al., 2014), the latter being characterized by a deepening of the oxic zone, and an increased input of dissolved gases and inorganic nutrients into the mixed layer (Sarmiento et al., 2006;

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Borges et al., 2011). Sampling was carried out in the Northern Basin (1.72° S, 29.23° E) in February 2012 (rainy season), and in the Northern Basin and Southern Basin (2.34° S, 28.98° E) in September 2012 (dry season).

O₂ concentration was measured with a YSI-proODO probe with a optical O₂ sensor (detection limit is 3 μmolL⁻¹), calibrated using air saturated water. Hereafter, “low-oxygen waters” stands for waters with concentration < 3 μmolL⁻¹. Lake water was collected with a 7 L Niskin bottle (Hydro-Bios) at a depth interval of 5 m from the lake surface to the top of the monimolimnion, at 80 m.

2.2 Chemical analyses

Samples for CH₄ concentrations were collected in 50 mL glass serum bottles from the Niskin bottle with a tube, left to overflow, poisoned with 100 μL of saturated HgCl₂ and sealed with butyl stoppers and aluminium caps. Concentrations of CH₄ were measured by headspace technique (Weiss, 1981) using gas chromatography with flame ionization detection (GC-FID, SRI 8610C), after creating a 20 mL headspace with N₂ in the glass serum bottles, and then analyzed following the method described by Borges et al. (2011). Samples for the determination of the δ¹³C signature of CH₄ (δ¹³C-CH₄) were collected in 250 mL glass serum bottles similarly to CH₄ concentration samples. δ¹³C-CH₄ was determined by a custom developed technique, whereby a helium headspace was first created, and CH₄ was flushed out through a double-hole needle, CO₂ was removed with a CO₂ trap (soda lime), and the CH₄ was converted to CO₂ in an online combustion column similar to that in an Elemental Analyzer (EA). The resulting CO₂ was subsequently preconcentrated by immersion of a stainless steel loop in liquid nitrogen in a custom-built cryofocussing device, passed through a micropacked GC column (HayeSep Q 2m, 0.75 mm ID ; Restek), and finally measured on a Thermo DeltaV Advantage isotope ratio mass spectrometer (IRMS). Certified reference standards (IAEA-CO1 and LSVEC) were used to calibrate δ¹³C-CH₄ data.

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Samples for the determination of $\delta^{13}\text{C}$ signatures of dissolved inorganic carbon (DIC) were collected by gently overfilling 12 mL glass vial (Labco Exetainer), preserved with 20 μL of HgCl_2 saturated. For the analysis of $\delta^{13}\text{C}$ -DIC, a 2 mL helium headspace was created and 100 μL of H_3PO_4 (99%) was added into each vial to convert all DIC species into CO_2 . After overnight equilibration, a variable volume of the headspace was injected into an EA coupled to an isotope ratio mass spectrometer (EA-IRMS; Thermo FlashHT with Thermo DeltaV Advantage). Calibration of $\delta^{13}\text{C}$ -DIC measurements was performed with certified reference materials (LSVEC and either NBS-19 or IAEA-CO-1).

Samples for particulate organic carbon (POC) concentrations and its stable C isotope signature ($\delta^{13}\text{C}$ -POC) were filtered on pre-combusted (overnight at 450°C) 25 mm glass fiber filters (Advantec GF-75; 0.3 μm), and dried. These filters were later decarbonated with HCl fumes for 4 h, dried and packed in silver cups. POC and $\delta^{13}\text{C}$ -POC were determined on an EA-IRMS (Thermo FlashHT with Thermo DeltaV Advantage). Calibration of POC and $\delta^{13}\text{C}$ -POC was performed with IAEA-C6 and acetanilide, and reproducibility of $\delta^{13}\text{C}$ -POC measurements was typically better than 0.2‰.

Samples ($\sim 2\text{ L}$) for measurements of phospholipid fatty acid concentrations (PLFA) and their $\delta^{13}\text{C}$ signature were filtered on pre-combusted 47 mm glass fiber filters (Advantec GF-75; 0.3 μm), and kept frozen until further processing. Extraction and derivatisation of PLFA was performed following a modified Bligh and Dyer extraction, silica column partitioning, and mild alkaline transmethylation as described by Boschker et al. (2004). Analyses were made on a Isolink GC-c-IRMS coupled to a Thermo DeltaV Advantage. All samples were analyzed in splitless mode, using an apolar GC column (Agilent DB-5) with a flow rate of 2 mL min^{-1} of helium as carrier gas. Initial oven temperature was set at 60°C for 1 min, then increased to 130°C at 40°C min^{-1} , and subsequently reached 250°C at a rate of 3°C min^{-1} . $\delta^{13}\text{C}$ -PLFA were corrected for the addition of the methyl group by a simple mass balance calculation, and were calibrated using internal (C19:0) and external (mixture of C14:0, C16:0, C18:0, C20:0, C22:0)

fatty acid methyl ester (FAME) standards. Reproducibility was estimated to be $\pm 0.6\%$ or better for natural abundance samples.

2.3 Determination of the isotope fractionation factor

In September 2012, the isotope fractionation factor (ϵ) was estimated by monitoring the changes in CH_4 concentration and $\delta^{13}\text{C}-\text{CH}_4$ over time in microcosms at several depths across the oxycline. Six glass serum bottles (60 mL) were gently overfilled at each depth and tightly capped with a butyl rubber stopper and an aluminium cap. They were then incubated in the dark at the lake temperature during 0, 24, 48, 72, 96 or 120 h. The incubation was stopped by poisoning the bottles with 100 μL of saturated HgCl_2 . The measurement of the concentration of CH_4 and the $\delta^{13}\text{C}-\text{CH}_4$ in every bottle was performed as described before. The isotope fractionation factor was calculated according to Coleman et al. (1981).

2.4 Methanotrophic bacterial production and growth efficiency measurement

At several depths throughout the water column, the methanotrophic bacterial production and methanotrophic bacterial growth efficiency were estimated by quantifying the incorporation of ^{13}C -labelled CH_4 ($^{13}\text{C}-\text{CH}_4$, 99.9%, Eurisotop) into the POC and DIC pool. Water from each sampling depth was transferred with a tube into 12 serum bottles (60 mL), capped with butyl stoppers and sealed with aluminium caps. Thereafter, 4 different volumes (50, 100, 150, or 200 μL) of a $^{13}\text{C}-\text{CH}_4$ gas mixture (1 : 10 in He) were injected in triplicate and 100 μL of saturated HgCl_2 was immediately added to one bottle per gas concentration treatment, serving as control bottle without biological activity. After vigorous shaking, the bottles were incubated in the dark during 24 h at the lake temperature. The incubation was stopped by filtration of a 40 mL subsample on 25 mm glass fiber filters (Advantec GF-75; 0.3 μm) to measure the ^{13}C -POC enrichment, and a 12 mL Exetainer was filled and poisoned with the addition of HgCl_2 in order to measure the ^{13}C -DIC enrichment. The exact amount of $^{13}\text{C}-\text{CH}_4$

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added in the bottles was determined from the bottles poisoned at the beginning of the experiment. The measurements of the concentration of POC, the $\delta^{13}\text{C}$ -POC, the $\delta^{13}\text{C}$ -DIC and the $\delta^{13}\text{C}$ - CH_4 were performed as described above. Methanotrophic bacterial production (MBP, $\mu\text{molL}^{-1} \text{d}^{-1}$) rates were calculated according to Hama et al. (1983):

$$\text{MBP} = \text{POC}_f \cdot \left(\%^{13}\text{C}\text{-POC}_f - \%^{13}\text{C}\text{-POC}_i \right) / \left(t \cdot \left(\%^{13}\text{C}\text{-CH}_4 - \%^{13}\text{C}\text{-POC}_i \right) \right) \quad (1)$$

where POC_f is the concentration of POC at the end of incubation (μmolL^{-1}), $\%^{13}\text{C}\text{-POC}_f$ and $\%^{13}\text{C}\text{-POC}_i$ are the percentage of ^{13}C in the POC and the end and the beginning of incubation, t is the incubation time (d^{-1}) and $\%^{13}\text{C}\text{-CH}_4$ is the percentage of ^{13}C in CH_4 directly after the inoculation of the bottles with the ^{13}C tracer.

The methanotrophic bacterial respiration rates (MBR, $\mu\text{molL}^{-1} \text{d}^{-1}$) were calculated according to:

$$\text{MBR} = \text{DIC}_f \cdot \left(\%^{13}\text{C}\text{-DIC}_f - \%^{13}\text{C}\text{-DIC}_i \right) / \left(t \cdot \left(\%^{13}\text{C}\text{-CH}_4 - \%^{13}\text{C}\text{-DIC}_i \right) \right) \quad (2)$$

where DIC_f is the concentration of DIC after the incubation (μmolL^{-1}), $\%^{13}\text{C}\text{-DIC}_f$ and $\%^{13}\text{C}\text{-DIC}_i$ are the final and initial percentage of ^{13}C in DIC. Finally, the methanotrophic bacterial growth efficiency (MBGE, %) was calculated according to:

$$\text{MBGE} = \text{MBP} / (\text{MBP} + \text{MBR}) \cdot 100 \quad (3)$$

The CH_4 concentration in the bottles sometimes increased drastically because of the $^{13}\text{C}\text{-CH}_4$ addition, which could have induced a bias in the estimation of MBP and MBR in case of CH_4 -limitation of the methanotrophic bacteria community. However, performing incubation along a gradient of CH_4 concentrations allowed us to assess if the measured MBP and MBR were positively related to the amount of tracer inoculated in the bottles. In case of such an effect (only at 50m in the Northern Basin in February 2012 and at 60m in the Southern Basin in September 2012) we applied a linear regression model ($r^2 > 0.90$) to estimate the intercept with the y axis, which was assumed to correspond to the MBP or MBR rates at in-situ CH_4 concentration.

2.5 Stable isotope probing of PLFA (SIP-PLFA) with $^{13}\text{C}\text{-CH}_4$

At each sampling depth and in parallel with the MBP measurement, 4 serum bottles (250 mL) were filled with water, overflowed and sealed with butyl stopper and aluminium caps. Bottles were spiked with 500 μL of $^{13}\text{C}\text{-CH}_4$ (99.9%). After 24 h of incubation in the dark at lake temperature, the water from the 4 bottles was combined and filtered on a single pre-combusted 47 mm glass fiber filter (Advantec GF-75; 0.3 μm) to quantify the incorporation of the tracer in bacterial PLFA. The filters were kept frozen until further processing. The extraction, derivatisation and analysis by GC-c-IRMS were carried out as described above.

3 Results

3.1 Physico-chemical parameters

In September 2012, the water column in the Southern Basin was oxic ($> 3 \mu\text{mol L}^{-1}$) from the surface to 65 m (Fig. 1a). CH_4 was abundant in deep waters, with a maximum concentration of 899 $\mu\text{mol L}^{-1}$ at 80 m, however CH_4 decreased abruptly in the oxycline (50–65 m), being 4 orders of magnitude lower in surface waters (Fig. 1a). Consistent with its biogenic origin, the upcoming CH_4 was depleted in ^{13}C in deep waters ($\delta^{13}\text{C}\text{-CH}_4$: -55.0‰) but became abruptly enriched in ^{13}C at the oxic–anoxic transition, where CH_4 concentrations sharply decreased, to reach a maximal value of -39.0‰ at 62.5 m depth (Fig. 1a). The $\delta^{13}\text{C}\text{-POC}$ values mirrored the pattern of $\delta^{13}\text{C}\text{-CH}_4$: they were almost constant from the surface to 55 m ($-24.4 \pm 0.3\text{‰}$), then showed an abrupt excursion towards more negative values at the bottom of the oxycline, with a minimum value (-42.8‰) at 65 m depth (Fig. 1a). Similar results were found in September 2012 in the Northern Basin, where the water was oxic ($> 3 \mu\text{mol L}^{-1}$) down to 55 m (Fig. 1b). At the oxic–anoxic transition, an abrupt isotopic enrichment of the

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CH₄ was also observed and the $\delta^{13}\text{C}$ -POC was relatively depleted in ^{13}C , similarly as in the Southern Basin (Fig. 1b).

In February 2012 in the Northern Basin, the water was oxic until 45 m depth (> 3 μmolL^{-1}) but the O₂ concentrations were below the limit of detection deeper in the water column (Fig. 1c). The gradual decrease in the CH₄ concentration between 60 and 45 m (from 110 to 3 μmolL^{-1}) was accompanied by a parallel increase of the $\delta^{13}\text{C}$ -CH₄ signature in the same depth interval (from -55.9 to -41.7‰), the residual CH₄ becoming isotopically enriched as CH₄ concentration decreased (Fig. 1c). $\delta^{13}\text{C}$ -POC values were also slightly lower below the oxic zone, with a minimum at 50 m (-26.9‰) (Fig. 1c).

3.2 Phospholipid fatty acid concentration and stable isotopic composition

Figure 2 shows profiles of the relative concentration and the $\delta^{13}\text{C}$ signature of specific PLFA in September 2012 (Fig. 2a and b; Southern basin) and February 2012 (Fig. 2c and d; Northern Basin). Irrespective of station, season and depth, the C16:0 saturated PLFA was always the most abundant PLFA (18–35% of all PLFA). The relative abundance of the C16 monounsaturated fatty acids (C16 MUFA) significantly increased at the bottom of the oxycline in February and September 2012. The $\delta^{13}\text{C}$ signature of the C16 MUFA was comparable to the $\delta^{13}\text{C}$ signature of the C16:0 in oxic waters, oscillating around -27 or -29‰ in February and September 2012, respectively. However, C16 MUFA were largely depleted in ^{13}C in the oxycline, with minimal $\delta^{13}\text{C}$ values as low as -55.3‰ at the oxic–anoxic transition in September 2012, and -49.5‰ in February 2012. This very strong depletion in $\delta^{13}\text{C}$ was only observed for this particular type of PLFA (C16 MUFA). The C18 MUFA were slightly more abundant in oxic waters (on average 9%) than in deeper waters (1–4%). Their isotopic composition varied with depth following the same vertical pattern than C16 MUFA, but with a lower amplitude. C18 MUFA minima in $\delta^{13}\text{C}$ were observed in low-oxygen waters in February 2012 (55 m, -35.1‰) and September 2012 (70 m, -30.5‰). The

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relative abundance of iso- and anteiso-branched C15:0 PLFA was systematically low (1–5%) and did not follow any depth pattern. Their isotopic signature was however slightly lower in oxygen-depleted waters than in oxic waters.

3.3 Isotope fractionation factor determination

5 During the isotope fractionation factor experiment, a significant decrease of the CH₄ concentration over time and a parallel enrichment of the residual CH₄ (Fig. 3) were monitored in every bottle incubated under oxic conditions. However, no consumption of CH₄ was measured in low-oxygen conditions. The isotope fractionation factor measured at several depths across the oxycline ranged between 1.008 and 1.024, and averaged 1.016 ± 0.007 (*n* = 5).

3.4 Methanotrophic bacterial production

MBP rates within the oxycline were variable (from 0 to 7.0 μmol C L⁻¹ d⁻¹). Maximum values were always observed near the oxic–anoxic transition (Fig. 1d–f), however substantial MBP (up to 2.2 μmol L⁻¹ d⁻¹) were also recorded under low-oxygen conditions (< 3 μmol L⁻¹) in February 2012 (Fig. 1f). Vertically integrated over the water column, MBP rates were estimated at 28.6 and 8.2 mmol m⁻² d⁻¹ in September 2012 in the Southern and Northern Basin, respectively, and 12.2 mmol m⁻² d⁻¹ in February 2012 in the Northern Basin. MBGE was found to be highly variable in the water column ranging between 50% at 52.5 m in the Northern Basin (September 2012) and 2% at 67.5 m in the Southern Basin (September 2012). Computed from depth-integrated MBP and MBR rates, the water column mean MBGE were 23% in September 2012 in the Southern and Northern Basins, and 42% in February 2012 in the Northern Basin.

Specific CH₄-derived C incorporation rates in PLFA (d⁻¹; incorporation rates normalized on PLFA concentration) show that bacteria containing C16 MUFA and C14:0 were particularly active in CH₄-derived C fixation in the oxycline in February

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and September 2012 (Fig. 4a and b). In contrast, the specific incorporation pattern was dominated by C17 MUFA, and to a lesser extent 10Me16:0 and C16 MUFA in low-oxygen waters in February 2012 (Fig. 4b).

4 Discussion

The sharp decrease of CH₄ concentration and the isotopic enrichment of the residual CH₄ in the oxycline, mirrored by the isotopic depletion of the POC pool at these depths indicated that microbial CH₄ oxidation is a strong CH₄ sink within the water column of Lake Kivu. Similar patterns characterized by a strong isotopic depletion of the POC pool in the oxycline were reported in other systems, such as the meromictic Northern Basin of Lake Lugano (Lehmann et al., 2004; Blees et al., 2014).

The fraction of the upward CH₄ flux oxidized within a depth interval can be estimated from a model of isotope fractionation for open systems at steady-state described by the following Rayleigh equation (Eq. 4; Bastviken et al., 2002):

$$f = \left(\delta^{13}\text{CH}_{4b} - \delta^{13}\text{CH}_{4t} \right) / \left(\left(\delta^{13}\text{CH}_{4t} + 1000 \right) \cdot \left((1/\alpha) - 1 \right) \right) \quad (4)$$

where f is the fraction of CH₄ oxidized within the depth interval, $\delta^{13}\text{CH}_{4b}$ and $\delta^{13}\text{CH}_{4t}$ are the $\delta^{13}\text{C}$ values of CH₄ at the bottom and the top of the depth interval, respectively, and α is the isotope fractionation factor for CH₄ oxidation estimated in Lake Kivu in September 2012 ($\alpha = 1.016 \pm 0.007$). Based on this equation and using a range of isotope fractionation factors (from 1.009 to 1.023), we can estimate that 57–105 % of the upward flux of CH₄ was microbially oxidized within a 10 m depth interval in the oxycline (60–70 m) in the Southern Basin during the dry season (September 2012). Similarly, 67–142 % of the CH₄ flux was oxidized between 50 and 55 m in the Northern Basin during the dry season, and 55–93 % of the CH₄ flux was oxidized within a wider depth interval (45–70 m) during the rainy season (February 2012). The relatively wide range of the estimated percentage of CH₄ flux oxidized is due to the uncertainty

on the isotope fractionation factor. Nevertheless, these calculations illustrate clearly the importance of microbial CH₄ oxidation processes in preventing CH₄ to reach the surface waters of the lake.

The theoretical δ¹³C signature of methanotrophs can be estimated at each depth from δ¹³C-CH₄ values and the experimental isotope fractionation factor (α, ranged between 1.009–1023). Then, applying a simple isotope mixing model with the δ¹³C signature of methanotrophs as an end-member and the the δ¹³C-POC in surface as a sedimenting organic matter end-member, it is possible to estimate the contribution of CH₄-derived C to the POC pool. Indeed, the contribution of CH₄-derived C appeared to be substantial at the bottom of the mixolimnion. In September 2012 in the Southern Basin, 32–44 % of the depth-integrated POC pool in the oxycline (between 60 and 70 m) originated from CH₄ incorporation, with a local maximum at the oxic–anoxic transition (65 m, 44–54 %). In the Northern Basin, 13–16 % of the POC in the oxycline (between 50 and 60 m) derived from CH₄. However, the contribution of CH₄ to the POC pool was relatively lower during the rainy season, as only 4–6 % of the POC in the 50–70 m depth interval, below the oxycline, had been fixed by methanotrophic organisms in the Northern Basin in February 2012 (local maximum slightly below the oxycline at 50 m, 8–10 %).

¹³CH₄ tracer experiments allowed to estimate the net MBP and the MBGE. Whatever the season, the highest MBP (0.8–7.2 μmol C L⁻¹ d⁻¹) rates were found in the oxycline. Hence, CH₄ oxidation in Lake Kivu seems to be mainly driven by oxic processes. Furthermore, maximal MBP rates were observed where the in situ CH₄ : O₂ ratio ranged between 0.1 and 10 (molar units, Fig. 5a), encompassing the stoichiometric CH₄ : O₂ ratio for aerobic microbial CH₄ oxidation (0.5) and the optimal ratio estimated in culture experiment (0.9, Amaral and Knowles, 1995). This relationship highlights the importance of the regulation of aerobic methanotrophic production by both CH₄ and O₂ availability. Vertically integrated over the water column, the MBP was estimated at 12.2 mmol m⁻² d⁻¹ during rainy season in the Northern Basin, and 28.6 and 8.2 mmol m⁻² d⁻¹ during dry season in the Southern Basin and the Northern Basin,

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respectively. These rates are comparable to the gross CH₄ oxidation rate reported earlier by Jannasch (1975) in Lake Kivu (7.2 mmol m⁻² d⁻¹) and the upward CH₄ flux recently estimated (9.38 mmol m⁻² d⁻¹) by Pasche et al. (2009). Areal MBP in Lake Kivu are equivalent to 16–58 % of the mean annual phytoplankton primary production (49 mmol m⁻² d⁻¹, Darchambeau et al., 2014), suggesting that biomass production by methanotrophs has the potential to sustain a significant fraction of the pelagic food-web. For example, it has been shown that cyclopoid copepods (mesozooplankton) of Lake Kivu escape visual predators by migrating below the euphotic zone, sometimes down to low-oxygen waters (Isumbisho et al., 2006), where they might feed on CH₄-derived C sources.

The relative contribution of MBP to the autochthonous production in Lake Kivu was distinctly higher than those reported in 3 Swedish lakes during summer, where MBP was equivalent to 0.3 and 7.0 % of the phytoplankton production (Bastviken et al., 2003). This was unrelated to the phytoplankton production rates in the Swedish lakes that ranged between 7 and 83 mmol m⁻² d⁻¹ and encompassed the average phytoplankton production value in Lake Kivu (49 mmol m⁻² d⁻¹). The MBP rates in the Swedish lakes (based on ¹⁴C incubations) were, however, distinctly lower than in Lake Kivu, ranging between 0.3 and 1.8 mmol m⁻² d⁻¹. This difference is probably related to the high CH₄ concentrations at the oxic–anoxic transition zone in Lake Kivu, as MBP peaked in the Swedish lakes at CH₄ concentrations < 100 μmol L⁻¹, while MBP peaked in Lake Kivu at CH₄ concentrations one to two orders of magnitude higher. Kankaala et al. (2013) reported seasonally resolved (for the ice-free period) MBP in five small (0.004–13.4 km²) boreal humic lakes (with dissolved organic C concentrations ranging between 7 and 24 mg CL⁻¹) in southern Finland. In these lakes phytoplankton production and MBP were highly variable, ranging between 5 and 50 mmol m⁻² d⁻¹ and < 0.2 mmol C m⁻² d⁻¹ and 41 mmol m⁻² d⁻¹, respectively. MBP was significantly higher in the two smallest lakes (0.004–0.008 km²), characterized by high CH₄ concentrations (< 750 μmol L⁻¹) and permanent anoxia throughout the year in bottom waters. Considering a MBGE of 25 %, their MBP estimates corresponded to a highly

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variable percentage of phytoplankton production, between 35 and 100% in the two smallest lakes, and between 0.4 and 5.0% in the three larger lakes (0.04–13.4 km²), and therefore they proposed that the relative contribution of methanotrophic bacteria to the total autotrophic production in a lake is related to its size (Kankaala et al., 2013).

5 However, the results reported for the large (2370 km²) Lake Kivu do not fit with this general pattern, probably because of the permanent and strong stratification of its water column that on one hand promotes a long residence time of deep waters and the accumulation of CH₄, and on the other hand leads to very slow upward diffusion of solutes, promoting the removal of CH₄ by bacterial oxidation as it diffuses to the surface.

10 The MBGE found during this study was variable (2–50%), but within the range of reported values in fresh waters (15–80%, King, 1992; 6–72%, Bastviken et al., 2003). MBGE was negatively related to the CH₄:O₂ ratio (Fig. 5b), i.e., a smaller fraction of the oxidized CH₄ was incorporated into the biomass at the bottom of the oxycline, where O₂ availability was relatively limited compared to CH₄. It has been recently suggested that under O₂-limiting conditions, methanotrophic bacteria are able to generate energy (adenosine triphosphate) by fermentation of formaldehyde (Kalyuzhnaya et al., 2013), the key intermediate in the oxidation of CH₄. This CH₄-based fermentation pathway would lead to the production of excreted organic acids (lactate, formate, ...) from CH₄-derived C instead of converting CH₄ into cellular biomass. If the metabolic abilities for this process are ubiquitous in methanotrophic organisms, it may potentially occur within the water column of Lake Kivu, at the bottom of the oxycline or in micro-oxic zone, as suggested by the low MBGE values found at high CH₄:O₂ molar ratio.

25 Almost all known aerobic methanotrophic bacteria are phylogenetically affiliated to Proteobacteria, belonging either to the *Alphaproteobacteria* (also referred to type I methanotrophs) or *Gammaproteobacteria* (type II methanotrophs) classes (Hanson and Hanson, 1996). The two distinct groups differ in some important physiological characteristics. Notably, they use different C fixation pathway (ribulose monophosphate

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for type I; the serine pathway for type II) and possess different patterns of PLFA. C16 MUFA are especially abundant in the type I methanotrophs while the type II methanotrophs contain mainly C18 MUFA (Le Bodelier et al., 2009). Therefore, the larger amount of ^{13}C -depleted C16 MUFA found in the oxycline and the strong labelling of C16 MUFA during the incubation with ^{13}C - CH_4 indicate that the aerobic methanotrophic community was dominated by type I methanotrophs in the water column during this study. In contrast, Type II methanotrophs did not appear to contribute much to the CH_4 oxidation in Lake Kivu, in good agreement with the results of Pasche et al. (2011). The dominance of type I over type II methanotrophs has been frequently reported in various stratified freshwater (Sundh et al., 1995; Brees et al., 2014) or marine environments (Schubert et al., 2006; Schmale et al., 2012), but this recurrent observation is still difficult to explain. In a recent review, Ho et al. (2013) attempted to classify several genus of methanotrophs according to their life strategies, using the competitor/stress tolerator/ruderal functional classification framework (Grime, 1977). Since type I methanotrophs dominate the active community in many environments and are known to respond rapidly to substrate availability, they classified them as competitors, or competitors-ruderals. In contrast, they proposed that type II members would be able to persist in the environment in a reversible state of reduced metabolic activity under non-favourable conditions, and thus classified them as stress tolerator, or stress tolerator-ruderal. Relatively large availability of CH_4 and O_2 (Figs. 1 and 5) at the oxic–anoxic transition of Lake Kivu is a favourable environment for the competitor-ruderal bacterial communities that could explain the dominance of type I methanotrophs over type II methanotrophs in this lake.

Also, a significant MBP rate ($1.3\ \mu\text{mol L}^{-1}\text{d}^{-1}$) was measured under low-oxygen conditions ($< 3\ \mu\text{mol L}^{-1}$) at 60 m during the rainy season (February 2012). Moreover, the PLFA labelling pattern was drastically different, with a more important specific ^{13}C incorporation into 10Me16:0 and C17 MUFA instead of the C16 MUFA, relative to their concentrations. This different labelling pattern suggests that a different population of methanotrophs was active in CH_4 oxidation deeper in the water column. Archaea

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lack ester-linked fatty acids in their membrane and are therefore undetectable in PLFA analysis. However 10Me16:0 and C17 MUFA are known to be especially abundant in sulphate-reducing bacteria (Macalady et al., 2000), the syntrophic partner of anaerobic CH₄ oxidizing archaea (Boetius et al., 2000). Hence, the specific labelling of 10Me16:0 and C17 MUFA under low-oxygen conditions could indicate that a fraction of the upward flux of CH₄ was oxidized syntrophically with SO₄²⁻ reduction during the rainy season, and might support the hypothesis that SO₄²⁻-reducing bacteria grow on CH₄-derived carbon source supplied by anaerobic methane oxidizers within the archaea/SO₄²⁻-reducers consortium, as already suggested by the results of an in vitro labelling (¹³CH₄) study (Blumenberg et al., 2005).

5 Conclusions

Lake Kivu ranks globally among the lakes with the lowest CH₄ emissions to the atmosphere (Borges et al., 2011), although the deep layers of the lake contain a huge amount of dissolved CH₄. This apparent paradox is linked to its strong meromictic nature that on one hand promotes a long residence time of deep waters and the accumulation of CH₄, and on the other hand leads to very slow upward diffusion of solutes, promoting the removal of CH₄ by bacterial oxidation as it diffuses to the surface. Our knowledge on bacterial CH₄ oxidation in Lake Kivu has been so far based on circumstantial evidence such as mass balance considerations (Borges et al., 2011; Pasche et al., 2011) and a few incubations carried out almost 40 years ago (Jannasch, 1975). Here, we provide conclusive evidences on the occurrence of CH₄ oxidation in the oxycline of Lake Kivu using stable isotopic characterisation of a suite of carbon pools (CH₄, POC, PLFA) as well as rate measurements (MBP). Vertically integrated MBP ranged between 8 and 29 mmol m⁻² d⁻¹, and was higher than previously reported in other lakes (Bastvinken et al., 2003; Kankaala et al., 2013). MBP was equivalent to 16–58% of the average annual phytoplankton primary production, a fraction distinctly

higher than previously reported in other lakes, usually < 10 % (Bastvinken et al., 2003; Kankaala et al., 2006). Hence, methanotrophic bacteria could potentially sustain a significant fraction of the pelagic food-web in this oligotrophic CH₄-rich lake.

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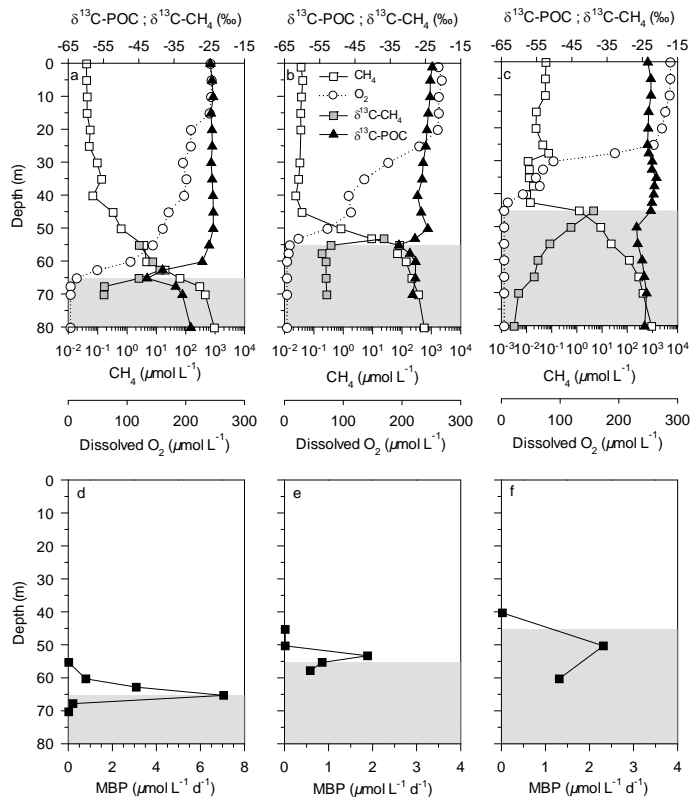
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Figure 1. Vertical profiles of dissolved O₂ concentration (μmolL⁻¹), CH₄ concentration (μmolL⁻¹), δ¹³C-CH₄ (‰) and δ¹³C-POC (‰) in Lake Kivu, in September 2012 (dry season) in the Southern Basin **(a)** and Northern Basin **(b)**, and in February 2012 (rainy season) in the Northern Basin **(c)**. Methanotrophic bacterial production rates (MBP, μmolL⁻¹d⁻¹) in September 2012 in the Southern Basin **(d)** and Northern Basin **(e)** and in February 2012 in the Northern Basin **(f)**. The grey zone corresponds to waters with dissolved O₂ concentration < 3 μmolL⁻¹.

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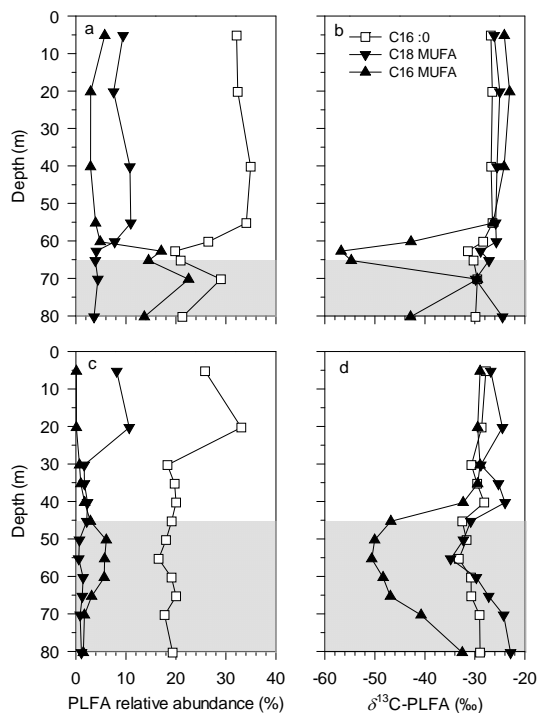


Figure 2. Vertical profiles of the relative abundance of phospholipid fatty acids (PLFA, %) and their respective carbon isotopic signature ($\delta^{13}\text{C-PLFA}$, ‰) in **(a, b)** the Southern Basin in September 2012 (dry season) and **(c, d)** in the Northern Basin in February 2012. The grey zone corresponds to waters with dissolved O₂ concentration < 3 μmol L⁻¹.

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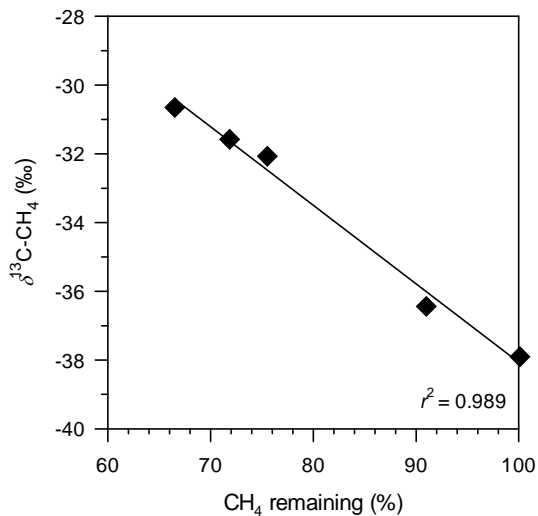


Figure 3. Example (62.5 m) of relationship between the $\delta^{13}\text{C-CH}_4$ and the fraction of CH_4 remaining in the bottles during the incubation to determine the isotope fractionation factor (%) carried out in September 2012 in the Southern Basin.

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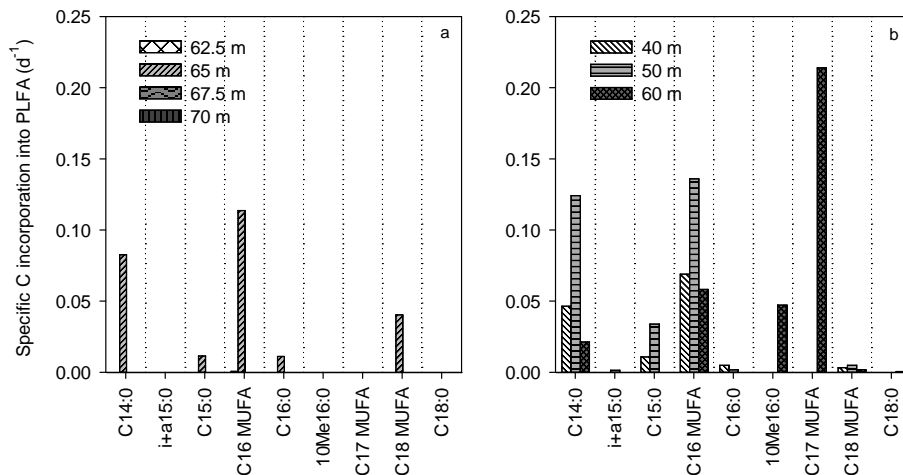


Figure 4. Specific CH_4 -derived C incorporation pattern into phospholipid fatty acids (PLFA) (incorporation rates of C into PLFA normalized on PLFA concentration, d^{-1}) in **(a)** September 2012 (dry season) in the Southern Basin and **(b)** in February 2012 (rainy season) in the Northern Basin.

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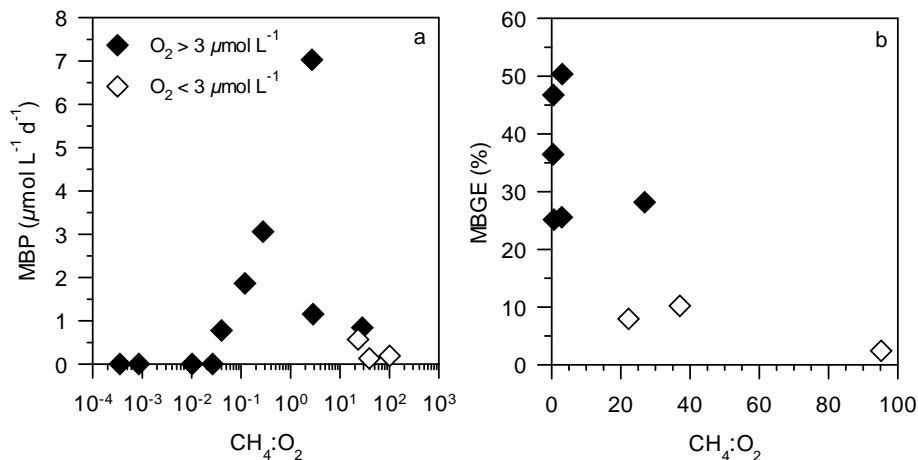


Figure 5. In Lake Kivu, **(a)** relationship between the methanotrophic bacterial production rates (MBP, $\mu\text{mol C L}^{-1} \text{d}^{-1}$) and the in situ $\text{CH}_4:\text{O}_2$ molar ratio and **(b)** relationship between the methanotrophic bacterial growth efficiency and the the in situ $\text{CH}_4:\text{O}_2$ molar ratio. The ratio was calculated with an O_2 concentration value of $3 \mu\text{mol L}^{-1}$ when observed in situ values were below the detection limit of the probe ($3 \mu\text{mol L}^{-1}$).