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CH_4 exchange at the forest floor of a forestry-drained fen: low flux rates but high temporal variation

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Abstract. We measured methane (CH_4) exchange rates with automatic chambers at the forest floor of a nutrient-rich drained peatland in 2011–2013. The fen, located in southern Finland, was drained for forestry in the 1970s and the tree stand is now a mixture of Scots pine, Norway spruce and pubescent birch. Our measurement system consisted of six transparent polycarbonate chambers and stainless steel frames, positioned on a number of different field and moss layer types. Flux rates were calculated with both linear and exponential regression. The use of linear regression systematically underestimated CH_4 flux rates by 20–50 % when compared to exponential regression. However, the use of exponential regression with small fluxes (< 3.5 µg CH_4 m^{-2} h^{-1}) typically resulted in anomalously large flux rates and high hour-to-hour deviations. We therefore recommend that flux rates are initially calculated with linear regression and that higher flux rates are then recalculated using exponential regression. There was clear seasonal dynamics in the CH4 flux: the forest floor acted as a CH_4 sink particularly from early summer until the end of the year, while in late winter the flux was very small and fluctuated around zero. However, the magnitude of fluxes was relatively small throughout the year, ranging mainly from –40 to + 50 µg CH_4 m^{-2} h^{-1}. CH_4 emission peaks were occasionally observed, particularly in spring during the snow melt, and in summer during heavy rainfall events. Diurnal variation was observed in all of the chambers, mainly in the summer and particularly in dry and warm conditions. Annual net CH_4 exchange among the six chambers varied from –31 to –155 mg CH_4 m^{-2} yr^{-1}; with an average of –67 mg CH_4 m^{-2} yr^{-1} over the two-year measurement period.

Keywords: methane uptake, methane emission, high resolution flux measurement, automatic chamber, peatland forest, drained peatland, linear regression, exponential regression

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1 Introduction

Methane (CH_4) is one of the important atmospheric trace gases due to its capability to absorb thermal radiation and warm the climate (IPCC, 2014). One of the main sources of CH_4 globally are peatlands (e.g. Denman et al., 2007) where CH_4 is produced by the decomposition of organic matter in anaerobic conditions (Waddington and Roulet, 2000; Rinne et al., 2007; Leppälä et al., 2011). Around 3 % of the Earth's land surface is covered by peatlands (4 000 000 km^2) (Clarke and Rieley, 2010) and the majority of peatlands are located in the boreal region (Fischlin et al., 2007). About one third (104 000 km^2) of European mire and peat resources are located in Finland (Joosten and Clarke, 2002; Montanarella et al., 2006) and more than half (55 000 km^2) of this area has been drained for forestry (Lappalainen, 1996; Päivänen and Hånell, 2012).

In peatlands, the net CH_4 flux between the soil and atmosphere is the sum of CH_4 production and oxidation (Dunfield et al., 1993). The drainage of peatlands results in water level drawdown and increased oxic layer thickness. Thereby, CH_4 production is decreased and the fraction of oxidized CH_4 increased (e.g. Moore & Knowles, 1989; Roulet et al., 1992). Consequently, the CH_4 oxidation rate in the aerated surface soil and mosses typically exceeds that of CH_4 production that occurs deeper in the soil, thus turning well-drained peatlands in particular into net CH_4 sinks (Martikainen et al., 1995; Minkkinen et al., 2007; Ojanen et al., 2010; Lohila et al., 2011). However, poorly drained sites may remain as CH_4 sources in many cases (Ojanen et al., 2010). In addition, the drainage ditches even at well-drained sites typically continue to emit CH_4 at rates similar to pristine boreal peatlands (Minkkinen et al., 1997; Minkkinen and Laine, 2006; Luan & Wu, 2015).

CH_4 is produced under anaerobic conditions by microbes known as methanogens. The rate of production is mainly controlled by soil temperature, pH and redox potential (Eh) (Dunfield et al., 1993; Wang et al., 1993; Kotsyurbenko et al., 2004). Low Eh values and a lack of electron acceptors other than acetate and hydrogen are a precondition for the production of CH_4 (Segers, 1998; Kotsyurbenko et al., 2004). Oxidation by methanotrophic microbes occurs in the oxic soil layer closer to the surface and also in the moss layer (Larmola et al., 2010). The rate is mainly controlled by oxygen concentration, temperature and soil moisture (Boeckx and Van Cleemput, 1996). In environments with low soil CH_4 production, uptake of atmospheric CH_4 by the methanotrophic microbes also takes place.

Closed chambers are commonly used in the measurement of greenhouse gas exchange between the forest floor and the atmosphere (e.g. Livingston and Hutchinson, 1995; Christensen et al., 1995; van Huissteden et al., 2005; Alm et al., 2007; Denmead, 2008; Forbrich et al., 2009, Koskinen et al., 2014). Unlike the eddy covariance (EC) method, which is more suitable for measuring fluxes at the ecosystem level, the chamber method permits the investigation of small scale processes, such as the gas exchange of different microtopographical surfaces, and enables the quantification of spatial variation (Keller et al., 1990; Singh et al., 1997). However, there are several details in the chamber structure and measurement technique that may have a significant impact on the observed flux rate, and hence should be taken into account. For example, increasing chamber size, especially height, seems to have an effect on the flux (Pihlatie et al., 2013). In addition, it has been suggested

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that chambers should include a fan to evenly distribute the air in the chamber headspace as it decreases uncertainty in the flux, although the rotational speed of the fan should be kept low to avoid excessive turbulence (Pumpanen et al., 2004; Christiansen et al., 2011, Koskinen et al., 2014). A second significant source of uncertainty is the impact of the chamber itself on the gas concentration gradient in the soil and in the boundary layer just above it (Healy et al., 1996; Davidson et al., 2002; Conen and Smith, 2000; Livingston et al., 2005). The concentration gradient is critical as it is the main factor driving the soil-atmosphere CH_4 exchange and disturbance may therefore have a severe impact on the observed flux.

The gradient between the soil and the air inside the chamber changes when the gas concentration inside the chamber changes during the measurement. This changes the flux rate, which makes the concentration change in time non-linear. There are many studies that have recognized that the use of linear regression in flux calculation can cause significant underestimation of the flux (e.g. Healy et al., 1996; Hutchinson et al., 2000; Pedersen et al., 2001; Welles et al., 2001; Nakano et al., 2004; Livingston et al., 2006; Kutzbach et al., 2007; Kroon et al., 2008; Pedersen et al., 2010; Pihlatie et al., 2010; Pihlatie et al., 2013) and that the error caused by using linear regression is systematic, not random (Hutchinson et al., 2000; Livingston et al., 2005; Livingston et al., 2006; Kutzbach et al., 2007). However, many studies have used linear regression (e.g. Reth et al., 2005; Laine et al., 2006; Wang et al., 2006; Alm et al., 2007; Jones et al., 2011; Bergier et al., 2013; Fassbinder et al., 2013), because under field conditions it is more robust than nonlinear methods. Moreover, the use of linear regression is preferred when comparing measurement sites as it is not as sensitive as non-linear models to small differences in soil properties (Venterea et al., 2009). The selection of the optimal fitting method is important as it can be the largest source of uncertainty in flux calculations (Levy et al., 2011; Venterea et al., 2013). However, the use of exponential regression is problematic as it is sensitive to disturbances during the measurement and it requires more than five measurement points. As such, a high resolution gas analyzer is typically required instead of traditional syringe sampling (Kroon et al., 2008).

In this study, we measured the CH_4 flux between the forest floor and atmosphere for 2 years in a relatively nutrient-rich forestry-drained peatland site with six automatic soil chambers and a cavity ring-down spectroscopy analyzer, which allowed us to sample with relatively high temporal resolution. Our particular aim with this set-up was to determine:

1. How large is the difference between fluxes calculated from linear and exponential regression, and which regression method should be preferred?
2. How large is the diurnal, seasonal or inter-variation in the CH_4 flux?
3. What is the annual CH_4 balance in the studied site?

2 Materials and methods

2.1 Site description

The measurements were made at Lettosuo in southern Finland (60°38' N, 23°57' E), a nutrient rich peatland forest that was drained and fertilized with phosphorus and potassium at the beginning of the 1970s. Before drainage, the tree stand was

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dominated by Scots pine (*Pinus sylvestris*) with some pubescent birch (*Betula pubescens*). After drainage, the site has become a mixture of Scots pine, pubescent birch, and Norway spruce (*Picea abies*) with understory vegetation. Stem volumes during 2011–2012 were estimated at 174, 46, and 28 m^3/ha for Scots pine, pubescent birch, and Norway spruce, respectively. The tree stand is quite dense and this results in irregular shading, which makes the ground vegetation layer patchy and variable. For example, herbs such as *Dryopteris carthusiana* and dwarf shrubs such as *Vaccinium myrtillus* are common in the ground vegetation (Koskinen et al., 2014). In addition, the moss layer is patchy and is dominated by *Pleurozium schreberi* and *Dicranum polysetum* with some *Sphagna (S. girgensohnii, S. angustifolium* and *S. russowii)* appearing in moist patches (Koskinen et al., 2014).

The climate at the site has both continental and maritime influences. The long-term (1981–2010) annual mean temperature and precipitation at the nearby weather station are 4.6 °C and 627 mm, respectively (Pirinen et al. 2012). Continuous EC measurements of carbon dioxide (CO_2) and latent and sensible heat fluxes have been running at the site since 2009. At the same time, measurements of meteorological variables, such as air temperature, relative humidity, global radiation, net radiation, photosynthetically active radiation (PAR), soil temperature profile, soil moisture, precipitation and soil heat flux have also been continuously recorded.

2.2 Measurement system

The measurements of forest floor CH_4 exchange started in March 2011 and continued until April 2013. CH_4 exchange was monitored with six transparent soil chambers placed at a distance of about 30 m from the EC mast (Koskinen et al., 2014). The locations of the chambers were selected so that they were close to each other but represented different ground vegetation types (Table 1).

The size of each chamber was 57 cm x 57 cm x 30 cm (l x w x h). A fan was used to mix the air inside the chamber headspace. A gas inlet tube made of polyurethane (FESTO, OD = 6 mm, ID = 4 mm) was positioned in the stream of the fan to mix the returning gas from the pipeline and prevent it returning to the outlet. Tubes (15 m in length) were used to transfer the gas to and from a measurement cabin where the mixing ratio of the gas was measured every four seconds with a Picarro G1130 cavity ring-down spectroscopy gas analyzer (Picarro Inc., Santa Clara, CA, USA).

The polycarbonate chamber was attached to a stainless steel frame (see description in Koskinen et al., 2014). The lower frame included five vertically movable legs, which could be used to keep the frame and chamber at snow level. A collar (height=5 cm) was used to connect the steel frame to the soil surface. Most of the roots in the area were left uncut as the collar was only installed to a depth of 2 cm in the moss layer. Peat and moss were used to seal the connection between the collar and soil. To prevent the gases from moving horizontally in the snow during winter, the steel frame was raised above the snow level by placing one or two extension collars (h=16 cm) between the frame and soil. When the chamber was closed, the connection between the frame (w=1 cm) and the chamber was sealed with silicone D-tape.

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Linear actuators (Linak Techline LA-35, Linak, 2009) were used to open and close the chambers. The flow rate of the sample gas in the system was kept at or slightly below 1 L min^{-1}. The tubes were flushed with ambient air just before the chamber closed. When all the chambers were open, ambient air was sampled.

Meteorological data was collected every 10 seconds from the chambers and adjacent areas with Nokeval 680-loggers. Temperatures inside the chambers were measured with pt100 temperature probes. The probe was installed at a height of 30 cm and positioned next to the fan under a metal heat shield to prevent direct solar radiation from affecting the measurements. Furthermore, soil surface temperatures were monitored inside each chamber just below the surface of the moss or litter layer. In addition, soil temperature probes were placed at depths of 2, 5, 10, 20 and 30 cm at one location near the chambers. Water table level (WTL) was monitored every hour from four different points at the site (TruTrack WT-HR data loggers, Intech Instruments Ltd).

2.3 Flux calculation

Each chamber was sampled usually once an hour with a closure time of 2, 5 or 8 minutes, with the exception of summer 2012 (JJA) when a longer closure time of 16 minutes was tested. Then, each chamber was sampled at bihourly intervals. For consistency between the measurements with different closure times, only the first 120 seconds of the concentration data from these measurements was used. However, data from longer closure times were utilized when testing the effect of different closure time (see 3.3).

Two different regressions types were fitted to the data: linear and exponential. The linear function describing the change in the concentration as a function of time [C(t)]was:

$$C(t) = a_{lin} + b_{lin} \times t, \quad (1)$$

where a_{lin} and b_{lin} are parameters and t is the time from the start of the closure. The concentration change in time is b_{lin} and the slope is assumed to be constant over the whole chamber closure.

The exponential function was:

$$C(t) = a_{exp} + b_{exp} \times \exp(c_{exp} t), \quad (2)$$

where a_{exp}, b_{exp} and c_{exp} are parameters. When differentiating Eq. (2) with time and inspecting the moment when the chamber closes (t=0), it follows that the concentration change with time is the product of parameters b and c. It is generally considered that this initial rate of concentration change best represents the flux rate of that time. However, when fitting the exponential function to the data using the least squares approach, the fitting usually fails due to local minima. To overcome this and to avoid overparameterisation, a second order polynomial is initially fitted to the data:

$$C(t) = a_2 t^2 + b_2 t + c_2, \quad (3)$$

where a_2, b_2 and c_2 are parameters. A Taylor power series expansion of the 17th order (Kutzbach et al., 2007) was fitted to the data to determine the initial estimate for the parameters of the exponential regression:

$$C(t) = a_{17} + b_{17} t + c_{17} t^2 + \sum_{i=3}^{17} \frac{2^{i-1} c_{17}^{i-1}}{i!\, b_{17}^{i-1}} t^i, \quad (4)$$

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where a_{17}, b_{17} and c_{17} are the initial estimates of parameters a_{exp}, b_{exp} and c_{exp} for Eq. (2).

The exponential regression should capture the flux better than the linear regression as it takes into account the change in the gradient between soil and chamber headspace during chamber closure, which is evident when diffusion flux is decreasing the concentration difference. However, exponential regression is very sensitive to possible disturbances at the beginning of chamber closure. In our study, we attempted to minimize these disturbances by closing the chamber slowly and smoothly, which seemed to prevent pressure peaks/fluctuations related to chamber closing.

The flux (F, μg CH_4 m^{-2} h^{-1}) was calculated according to the Eq. 5, which is based on the ideal gas law:

$$F = \left(\frac{dC(t)}{dt}\right)_{t=0} \cdot M \cdot \frac{P}{R} \cdot \frac{273.15}{273.15\,°C+T} \cdot \frac{V}{A} \cdot 3600\,\frac{s}{h}, \quad (5)$$

where $\left(\frac{dC(t)}{dt}\right)_{t=0}$ is the time derivative (ppm/h) of a linear (b_{lin}) or exponential ($b_{exp} \times c_{exp}$) regression at the beginning of the closure, M is the molecular mass of CH_4 (16.042×10^6 μg mol^{-1}), P is the air pressure (Pa), R is the universal gas constant (8.31446 J mol^{-1} K^{-1}), T is the mean chamber headspace temperature during closure (°C), and V and A are the volume (m^3) and the base area (m^2) of the chamber headspace, respectively. Air pressure, precipitation and snow depth data were acquired from the nearby Finnish Meteorological Institute observatory at Jokioinen (~35 km northwest of Lettosuo). Here, a micrometeorological sign convention is used: a positive flux indicates a flux from the ecosystem to the atmosphere (emission) and a negative flux indicates a flux from the atmosphere into the ecosystem (uptake).

When estimating the volume of the chamber headspace, the height of the moss and snow surfaces were assumed to represent the interface between the soil and air. In other words, the pore space in the soil and snow was ignored from the headspace volume. The error caused by this in flux calculations is estimated to be only a few percent (Koskinen et al., 2014). To create a continuous data set of snow depth, the manual measurements carried out irregularly at the site were combined with those measured daily at the Jokioinen observatory. In addition to snow depth, the height of the chamber headspace was measured at the start and end of the growing season from 16 points inside each collar by gently placing the end of a tape measure on top of the surface mosses (Koskinen et al., 2014). The height of the chamber headspace between these manual measurements was determined with linear interpolation.

2.4 Filtering of the flux data

After the fluxes were calculated, several filters were applied to remove cases when the measurement system did not work adequately. The most common reason for discarding data were problems associated with the hardware, for example, the improper functioning of a linear actuator, which caused the chambers to remain either stuck open or closed. These cases were detected by monitoring the simultaneously measured CO_2 concentration data during the closure. Goodness of fit was checked by calculating the normalized root mean square error (NRMSE) (e.g. Christiansen et al., 2011; Pihlatie et al., 2013) for each fit using Eq. 6:

$$NRMSE = \frac{\sqrt{\frac{1}{n}\sum_{i=1}^{n}(C_{fit,i}-C_i)^2}}{C_{max}-C_{min}}, \quad (6)$$

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where n is the number of measurement points, $C_{fit,i}$ is the concentration calculated from the fit, C_i is the measured concentration and C_{max} and C_{min} are the highest and lowest concentrations measured during closure. The numerator is also known as the root mean square error (RMSE). If the NRMSE was larger than 0.05, the CH_4 data from that closure was discarded. It should be noted that the application of this criteria removes closures with $\Delta CO_2 = 0$, which may result when photosynthesis equals respiration. Here we found < 20 of such cases meaning that this criteria could be applied without removing a significant amount of suitable data.

In addition to NRMSE filtering, a running mean of CH_4 flux (F_{CH4}) with a time window of 14 days (shifting one day at a time) and standard deviation (σ) was calculated to remove random spiking in the data. The data points that failed to fall within F_{CH4}±10σ were removed iteratively. In total, 75923 closures were recorded from which 19 % (n=14345) were discarded due to large NRMSE values (hardware problems) and 7 % (n=4999) with σ-filter.

Cumulative CH_4 flux was calculated by summing all the hourly measured fluxes for each chamber over one year of measurements. Possible gaps were filled with linear interpolation. As our measurements started in April 2011 and ended in March 2013, exactly two years, we decided to call the time period 4/2011–3/2012 the first year and 4/2012–3/2013 the second year.

A 95 % confidence interval was used to show the uncertainty of the results unless otherwise specified.

3 Results

3.1 Meteorological conditions

The first measurement year was significantly warmer (annual mean temperature 5.8 °C) than the second measurement year (1.4 °C) (Fig. 1). The first year was slightly warmer and the second year was significantly colder than the long-term mean recorded at the nearby weather station (4.6 °C). Both the summer (17.6 °C) and winter (2.7 °C) temperatures in 2011 were warmer than those of 2012 (12.1 °C and –2.3 °C). In particular, the beginning of summer 2012 was much cooler than for the same period in 2011.

Annual precipitation during the first (976 mm) and second (780 mm) measurement years was higher than the long-term mean (627 mm). Summertime precipitation was 9% higher in the first (309 mm) year as compared to the second (284 mm) year, while in winter the difference was 18% (577 mm and 490 mm in the first and second winters, respectively). The first snow appeared on 5th December 2011 and 25th October 2012, and the first permanent snow was recorded on 7th January 2012 and 28th November 2012 in the first and second measurement years, respectively. In the first spring, the snow had melted by 13th April 2011. For spring 2012, we do not know the exact day of snow melt due to missing data, although the snow had melted by 4th April. From the temperature data, we would estimate that the snow cover had disappeared sometime in mid-March.

WTL varied from –8 to –59 cm from soil surface and was highest in the spring and in late autumn. Lowest (i.e. deepest) values were reached at the end of summer. The average WTL in summer 2011 was –47.2±7.4 cm (±Standard deviation) and

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–49.1±7.1 (± Standard deviation) in summer 2012. Occasional sudden increases in WTL were observed after rainfall events and usually took 1–2 weeks to reach the WTL prior to the event.

3.2 Flux calculation method

The CH_4 fluxes calculated with linear regression (Eq. 1) were systematically and significantly lower than those calculated with exponential regression (Eq. 2) (Table 2a). At very low flux rates, the exponential regression resulted in increased variation in fluxes due to decreased signal-to-noise ratio in the gas mixing ratio time series. Typically, exponential regression created a sharp slope at the beginning of the time series that resulted in high variation in fluxes. We found that, on average, when the linear CH_4 flux values were below 3.5 µg CH_4 m^{-2} h^{-1} (Fig. 2), the use of the exponential fit resulted in noisy flux values. When the fluxes exceeded this value, linear regression resulted in an underestimation of the flux rates. Therefore, we decided to use the flux values estimated with the linear regression method when that method estimated fluxes lower than 3.5 µg CH_4 m^{-2} h^{-1}. If the flux was higher, we used the flux calculated from the exponential regression. Unless otherwise specified, the fluxes shown hereon have been calculated with this method.

When the data below and above the limit of 3.5 µg CH_4 m^{-2} h^{-1} was examined, the difference between the linear and exponential fits mainly varied between 20–50 % (average 35.3±0.3 %). The relative difference was dependent on the time of the year: it was largest during the winter (DJF) and spring (MAM) when the temperature and the soil CH_4 sink were at their lowest, and smallest in summer and autumn, varying from 19.0 to 23.8 %. For fluxes > 3.5 µg CH_4 m^{-2} h^{-1}, i.e. when the exponential regression did not work reliably, the overall difference between the linear and exponential fits decreased from 35.3±0.3 % to 22.1±0.2 % (Table 2b). Seasonally, the greatest effect was seen in the winter and spring data, as a large amount of data with fluxes below the limit were removed from this period.

The underestimation (in %) caused by using the linear regression was slightly smaller in 2012 compared to 2011. However, the uncertainties associated with the fluxes were slightly larger in 2012. Comparison of the fitting methods for the winter periods was meaningless as the exponential regression did not give reliable results in winter 2012–2013 as almost all the fluxes were < 3.5 µg CH_4 m^{-2} h^{-1}.

3.3 Effect of closure time on fluxes

No significant differences between the fluxes calculated with 2- or 5-minute closure times were detected. When the whole measurement period was examined, the average relative difference between the 2- and 5-minute closure times was 4.2±0.1 %, with the 2-minute closure always producing larger fluxes. However, the use of only linear regression to calculate the fluxes produced a larger difference for the whole measurement period (7.4±0.1 %). The differences were also checked annually and seasonally. The second year had a smaller difference (1.9±0.1 %) between the closure times than the first year (9.1±0.2 %), although it should be noted that in the first year most of the measurements were made with the 2-minute closure time and only 4 measurements per day had 5-minute closures. As such, there were fewer measurements available for the

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analysis for the first year. Seasonal comparisons revealed that the differences between the closure times were smallest in autumn and summer when the fluxes were largest, slightly larger in winter and largest in spring.

3.4 Seasonal dynamics of CH_4 flux and comparison between the chambers

During the 2-year measurement period, CH_4 flux rates varied mainly between –40 and +50 µg CH_4 m^{-2} h^{-1} (Fig. 3 & Fig. 4). Higher rates of uptake were measured more often during the first year (Fig. 4a) and during summer 2011 (Fig. 4b) than in the second year and in summer 2012, although the number of cases indicating emissions were low in both. Although the soil acted as a CH_4 sink for most of the time in all chambers, a few emission peaks of up to 90 µg CH_4 m^{-2} h^{-1} were recorded during and following heavy rainfall events. The filtering (see 2.5) used for the data was deliberately flexible to prevent the removal of these short-lasting CH_4 bursts from the soil. While some emission peaks were observable in all of the chambers, the peaks were largest in chamber #6. However, the peaks did not necessarily occur at the same time in different chambers.

CH_4 fluxes showed a clear seasonal variation. In spring, when the snow melted and thawing of the soil surface had started, CH_4 emissions fluctuating around 3 µg CH_4 m^{-2} h^{-1} were observed in all chambers (Fig. 3). As the temperature rise continued, chambers showed increasing CH_4 uptake. In both years, CH_4 uptake was largest in August, with fluxes varying between –15 and –40 µg CH_4 m^{-2} h^{-1}. In September, the uptake decreased and by November had dropped to half of that observed in the summer. However, the soil acted as a sink until the soil surface froze, after which CH_4 fluxes fluctuated around zero.

Considerable and systematic differences in fluxes between the chambers were detected. The largest two-year average values were measured in chambers #5 and #2 (Fig. 5), which were dominated by *Pleurozium schreberi* and *Dicranum polysetum*. The third largest sink was observed in chamber #6 (Fig. 5) with a *Sphagnum girgensohnii* carpet, while the remainder of the chambers showed similar flux rates. Annual net CH_4 exchange rates were on average –84±33 and –51±18 mg CH_4 m^{-2} yr^{-1} in 2011–2012 and 2012–2013, respectively. However, the forest floor sink was 21–54 % lower during the second annual period compared to the first. The largest year-to-year drop in annual CH_4 exchange was observed in the two chambers dominated by *Pleurozium schreberi* and *Dicranum polysetum* (#5 and #2), followed by the chamber dominated by *Sphagnum girgensohnii* (#6).

In four of the chambers (#2, #3, #5 and #6), a statistically significantly (Student's t-test, $p<0.05$) larger sink was observed in summer 2011 than in summer 2012. Moreover, the sink period, i.e. the period when all chambers acted as CH_4 sinks, was longer in the first than in the second year. In the first year, all chambers acted as sinks from 9th June 2011 to 31st January 2012, a total of 236 days. During the second year, the sink period lasted for 191 days from 22nd May 2012 until 29th November 2012. Among some chambers, a difference of six weeks was detected in the length of the sink period between years. The difference between the mean winter and summer fluxes was about 5–10 µg CH_4 m^{-2} h^{-1}; larger in 2011 than in 2012.

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3.5 Factors controlling the short- and long-term variations in CH_4 exchange

There was an observable, exponential relationship between CH_4 fluxes and the deeper soil temperatures when the spring and summer data of 2011 were pooled (Fig. 6a). Splitting this data into shorter periods showed that the relationship was rather strong in April–May and in June–start of July, but after that, when the soil temperature at 30 cm depth exceeded 12°C, the relationship was rather weak. Since it is likely that there is a co-correlation between soil temperature and WTL with higher CH_4 uptake at lower WTL (Fig. 6b), we plotted the residuals of Fig. 6a against WTL (Fig. 6c). This plot suggests that the high CH_4 uptake during the latter half of July and August cannot be solely explained by higher temperatures, or by the lower WTL. This increase of residuals usually started when the WTL was between –35 and –40 cm and when the WTL decreased even more, the oxidation was larger than acquired from the exponential model (Fig. 6a). All the chambers recorded similar behavior when 2011 and 2012 were compared, although in 2012 the data was noisier and this phenomenon was not as clear.

In addition to seasonal dynamics, diurnal variation in CH_4 flux rates was observed at least occasionally in all chambers. Most of this short-term variation was observed in summer (Fig. 7), with a few cases in spring 2011. At the beginning of June 2011 in particular, a strong diurnal cycle that coincided with the variation in air temperature was observed: Relatively high CH_4 uptake was observed during the morning hours (Figs. 7 and 8a) whereas the sink strength decreased towards afternoon. This period was associated with high daytime air temperatures and relatively low WTL (Fig. 7). However, during the latter half of June 2011 when there was lower ambient air temperatures and higher WTL, the diurnal variation was absent (Fig. 8b). However, in summer 2012 the largest uptake occurred in the afternoon and uptake was lowest at night (Fig. 8c). Similarly to the previous summer, the periods with diurnal variation had relatively low WTL, although daytime temperatures were generally 5–10 °C lower.

To understand the driving factors behind the diurnal and seasonal variation, correlations (Pearson) between the hourly CH_4 flux and environmental variables were checked separately for each chamber using monthly, seasonal and annual data. Only in the cases of insignificant correlations (usually $|r|<0.1$), the p-value reached values higher than 0.01. In some cases, a number of chambers showed positive and some chambers negative correlations of similar strength during the same season, for example, in the case of surface temperatures in spring and autumn 2011. The correlation coefficients in Table 3 have been classified accordingly.

There was a highly significant correlation between CH_4 flux and soil temperature. In general, soil temperatures at 20 cm (T20) and 30 cm (T30) depths correlated negatively with CH_4 flux rates (i.e., higher uptake at higher temperatures) in each of the eight seasons during the two study years. In addition, soil temperatures at 5 cm (T5) and 2 cm (T2) depths often correlated with CH_4 flux, however a marked difference was observed between summers 2011 and 2012: in 2011 the most significant correlation was observed with the deeper soil temperatures (T20 & T30) and with friction velocity (u^*), in 2012 the air and surface soil temperature, as well as T2 and T5 showed the best correlation with the flux. Interestingly, the soil temperature at 10 cm (T10) had the poorest correlation with the CH_4 flux, with significant correlation in all the chambers observed in only two seasons.

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There was a positive correlation between WTL and CH_4 flux in both summers (i.e. the deeper the WTL, the higher the uptake). A negative correlation with CH_4 flux, (i.e. the lower the WTL, the higher the emission) was only observed in winter 2011–2012. The correlation between the flux rate and PAR was always low or absent.

A significant (positive) correlation between u^* and the flux rate was only found in summer and autumn 2011. However, apportioning the flux data into u^* bins (interval 0.05 ms^{-1}) resulted in a steady CH_4 exchange rate with u^* of <0.25 ms^{-1} with a decreasing rate after that value (Fig. 9). This effect was also observed in shorter time periods, for example, in summer and autumn, although the results were not always significant. However, lengthening the inspected time period increased the statistical significance. Even though u^* was found to correlate with CH_4 flux when diurnal variation in the flux was observed (Fig. 8), we were not able to explain the variation with u* due to simultaneous correlation with ambient, soil surface, T2 and T5 temperatures.

4 Discussion

4.1 Impact of closure time and calculation method on fluxes

In this study, we found that using the linear regression method in flux calculations resulted in 20–50 % lower flux rate values for most of the time, in comparison to flux rates calculated with an exponential fit. In contrast, in winter and early spring, i.e., at the time of low CH_4 fluxes (< 3.5 µg CH_4 m^{-2} h^{-1}), linear regression gave more reliable results with lower noise. The uncertainty associated with exponential regression with low fluxes was caused by the decreased signal-to-noise ratio in the concentration data, leading to more or less arbitrary values of the concentration change over time at t=0. This unreliability explains why the difference between fluxes estimated by the linear and exponential was highest during winter and spring.

The use of exponential regression may be especially justified during summer and autumn when distortion of the vertical concentration gradient occurs inside the chamber. In this period, linear regression significantly underestimated the flux by an average by 21.5±1.1 %. In winter and spring, the underestimation was as high as 60 %. The mean underestimation during the whole measurement period (35.3±0.3 %) is in agreement with many previous studies, for example, Anthony et al. (1995) and Pedersen et al. (2010) reported a 35 % and 34 % decrease in flux values when using linear regression instead of exponential regression. Kutzbach et al. (2007) noted that underestimations varied mostly between 20–60 % on multiple sites in their study, while Pihlatie et al. (2013) under laboratory conditions at different flux levels reported an average underestimation of 30 %. For clarification, the distortion of the vertical concentration gradient in this case describes the effect where the concentration difference between the soil and the chamber headspace decreases during chamber closure, which causes the flux to change over time. This occurs every time that a vertical concentration gradient exists between the chamber headspace and the soil. In addition, the microbial oxidation rate is proportional to the CH_4 concentration in the soil (e.g. Ridgwell et al., 1999), which may have an effect on CH_4 concentrations within the chamber headspace.

Therefore, we recommend that in the future studies with temporally high-resolution data, the fluxes should be calculated initially with both methods to determine the threshold for "low" fluxes. As this threshold value is always dependent on the

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measurement system and method, it cannot be generalized. We are not aware of any publications where this would have been tested.

We also found that the length of the closure time had a small effect on the calculated CH_4 flux; the mean difference in the flux measured with the 2- or 5-minute closure times was 4.2±0.1 %. Koskinen et al. (2014) also tested how the length of the fit might affect CO_2 flux and found that with the linear fit the mean flux did not change significantly, although the RMSE and non-linearity increased. However, in their study the first 120 seconds of the closure were always removed before the start of fitting, and it is likely that using all the data would have resulted in significant differences between the different closure times. In this study, we noticed that the difference between the closure times was larger when using only linear regression to calculate the fluxes (7.4±0.1 %), which further supports the observation of increasing non-linearity with longer closure time. In addition, we noticed that the relative difference between the 2- or 5-minute closures was largest in winter and spring when CH_4 fluxes were lowest. Even though absolute differences in fluxes were slight, the small non-linearity in low fluxes, which became apparent over longer closure times, caused higher relative differences between the different dataset lengths in winter and spring. It is also possible that in the case of low fluxes, the use of longer closure times may be needed to increase the concentration change in chamber headspace to make the measurement less sensitive to the instrument noise. However, it should be noted that an increased closure time makes the measurements more prone to disturbance by wind gusts, especially if the chamber has a vent tube or if the measurements are made on soils with high porosity (Bain et al., 2005).

4.2 CH_4 exchange dynamics in a peatland forest

The measurement site was a small annual CH_4 sink (50–80 mg CH_4 m^{-2} yr^{-1}) over the two year measurement period. While we do not have measurements prior to drainage, the site was originally a herb-rich tall sedge pine fen and is similar to a site reported by Nykänen et al. (1998) with high CH_4 emissions (25 g CH_4 m^{-2} yr^{-1}). Therefore, assuming similar emissions at Lettosuo before drainage, the peat soil has obviously turned from a CH_4 source to a small CH_4 sink. Other studies in drained peatland forests have reported uptake that varied between 10 and 970 mg CH_4 m^{-2} yr^{-1} (Alm et al., 1999; Minkkinen et al., 2007; Ojanen et al., 2010; Lohila et al., 2011), while the average uptake in boreal upland forests is about 200 mg CH_4 m^{-2} yr^{-1} (Dutaur and Verchot, 2007). Thus, the CH_4 sink of Lettosuo soil was smaller than the average CH_4 sink in boreal upland forests. However, it should be noted that our calculations of annual and daily CH_4 exchange do not include the emissions from the ditches, which have been found to be highly variable: from 0 to 600 mg CH_4 m^{-2} d^{-1} (e.g. Minkkinen et al., 1997; Minkkinen and Laine, 2006; Luan & Wu, 2015). At Lettosuo, where the ditches cover 2–3 % of the area, we estimate that CH_4 emissions (m^2 peatland) from the ditches averaged (±standard error of the mean) 24 (±58) g CH_4 m^{-2} yr^{-1} (unpublished data). This would suggest that even with a very small ditch area, Lettosuo is a small annual source of CH_4 to the atmosphere, although the very high uncertainties associated with this calculation should be noted.

Many drained peatland sites have been shown to act as CH_4 sources over a year due to poor drainage and high water table levels (e.g. Nykänen et al., 1998; Minkkinen and Laine, 2006; Ojanen et al., 2010). At our site, small CH_4 emissions (< 30

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μg CH_4 m^{-2} h^{-1}) were recorded during winter and spring when the soil temperature was close to zero. These low temperatures are likely to slow down the activity of methanotrophic bacteria (e.g. Boeckx and Van Cleemput, 1996) and low temperatures also decrease the rate of microbial CH_4 production (e.g. Dunfield et al., 1993). In addition, a few larger CH_4 emission peaks were observed during and after heavy rainfall events, but not all chambers responded to the same rainfall events. The largest number and magnitude of these short bursts of CH_4 were recorded with the *Sphagnum*-dominated chamber (#6), which was expected as *Sphagnum* mosses favor wet spots. These rainfall events turned the soil from a sink to a small source for a short period, possibly due to increased water saturation and decreased oxic space, which could promote methanogenic activity. Nykänen et al. (1995) also observed that a soil can switch to a CH_4 source (up to 0.2 mg CH_4 m^{-2} h^{-1}) during increased water saturation. However, the emission peaks at Lettosuo were relatively small when compared to some upland mineral forest sites, where emissions of up to 3.7 mg CH_4 m^{-2} h^{-1} have been observed in wet conditions (e.g. Savage et al., 1997; Lohila et al., 2016). Similarly, the maximum hourly uptake in summer at Lettosuo was slightly lower than reported for the above mentioned upland forests (on average 20 μg CH_4 m^{-2} h^{-1} at Lettosuo vs. 40–80 μg CH_4 m^{-2} h^{-1} at the upland forests).

There were relatively large differences in the net CH_4 exchange rates between the chambers (largest uptake 154 mg CH_4 m^{-2} yr^{-1}, smallest uptake 32 mg CH_4 m^{-2} yr^{-1}), even though all the chambers were only a few meters apart from each other. In our measurements, the difference in the soil surface temperature between the chambers was usually less than 2 degrees, which would indicate that the soil temperature was not the main factor determining the observed spatial variation in fluxes. Moreover, since we do not have WTL data for below each of the chambers, we cannot confirm its role in explaining the difference. However, it is unlikely that the variation in the WTL solely could explain the difference: chambers #4 and #5 were located closest to the ditch, and probably had the deepest WTL. Indeed, chamber #5 showed the highest uptake, although chamber #4 was one of the smallest sinks. Hence, although WTL is likely to explain part of the variation in the spatial variation of the CH_4 flux, there are potentially many other factors, such as the vegetation type or small-scale soil properties.

All the chambers recorded diurnal variation in CH_4 flux at some time in the study with most of the variation observed during summer. In the beginning of June 2011, uptake was at its highest during the early morning, while in June 2012 it was highest in the afternoon (Figs. 7 and 8). The first half of June 2011 was dry and warm, whereas June 2012 was cooler (Fig. 8). In the first half of June 2011 the daytime temperatures reached as high as 30 °C and a strong diurnal pattern was observed with the highest uptake values observed early in the morning and the lowest in the afternoon. This diurnal variation ceased after mid-June 2011 with increased rainfall and decreased temperatures. In the beginning of June 2011, CH_4 oxidation in the afternoon might have been hindered due to high daytime temperatures, which likely resulted in a drier surface soil layer. In contrast, drying of the soil surface did not occur in summer 2012, and the CH_4 oxidation peaked in afternoon simultaneously with the soil temperature maximum. The cycle in summer 2012 had a similar pattern in regard to the highest and lowest uptake rates as has also been observed by Wang et al. (2013) in a Canadian mixed forest. In CH_4 emitting ecosystems, such as natural and

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experimental wetlands and in lakes, CH_4 emissions have been reported to peak in afternoon and reach their minimum at night (e.g. Mikkelä et al., 1995; Duan et al., 2005; Morin et al., 2014, Sun et al., 2014).

During periods of a discernable diurnal cycle, the daily range of observed fluxes could vary by as much as 20 μg CH_4 m^{-2} h^{-1}. This would suggest that determining CH_4 exchange for longer periods from only daytime measurements using, for example, manual chambers, could cause a significant bias in annual balance calculations. However, when the flux is correlated with temperature, the flux can be modeled with temperature as an explanatory variable to decrease the bias (e.g. Ojanen et al., 2010). If measurements take place only during the growing season and these are then used to estimate annual CH_4 exchange it may result in an overestimation of the sink. We tested this by taking one daytime measurement from each chamber every second week, allowing a few days and a few hours of random variation around this two week time step. This provided a selection of multiple datasets for the same time period. Annual CH_4 exchange was calculated using linear interpolation between the points, as is often done when using manual chamber data. The resulting values were compared to CH_4 exchange calculated from our automatic chamber measurements (Fig. 3). The decreased measurement interval underestimated CH_4 uptake by 11–54 % (average 30±3 %) depending on the chamber.

In this study, CH_4 flux correlated best with the soil temperatures measured at the depths of 20 cm and 30 cm, and WTL, but also with air, soil surface and soil temperatures at depths of 2 cm and 5 cm during summer and autumn. CH_4 fluxes (both emission and uptake) have been found to correlate with WTL in many studies (Mikkelä et al., 1995; Bellisario et al., 1999; Ojanen et al., 2010; Mastepanov et al., 2013). Here we found evidence that WTL controls the CH_4 flux, particularly when WTL < –40 cm, showing increased CH_4 uptake rates with deeper WTL. Due to the likely co-correlation of WTL and soil temperature, both typically peaking in July, determination of the exact contribution of these variables on CH_4 flux is not straightforward. However, from the analysis shown in Figs. 6a-c it seems likely that the influence of WTL in controlling the CH_4 exchange is more pronounced at deeper WTL. First, this might be attributed to a decrease in CH_4 emissions from the deeper layers; a logical consequence of the drawdown of the anoxic peat layer suitable for CH_4 production. On the other hand, increased net CH_4 uptake could be caused by increased CH_4 oxidation due to the increase of the oxic peat layer. The observed negative correlation between the flux rate and soil temperature suggests increased consumption of CH_4 by methanotrophs as it is known that increasing soil temperature enhances methanotrophic activity (e.g. van den Pol-van Dasselaar et al., 1998; Mohanty et al., 2007). In other studies, significant correlations between CH_4 flux and temperature have typically been found in ecosystems that display CH_4 emissions (Nykänen et al., 1998, Minkkinen & Laine, 2006), although the direction of the correlation has been found to differ in fens and bogs, respectively. In contrast, no significant correlations with temperature have been found in peatland forests that mainly show CH_4 uptake (Ojanen et al., 2010; Wang et al., 2013). In pristine peatlands, temperature has been shown to correlate positively with the CH_4 emission rate (e.g. Mikkelä et al., 1995; Bellisario et al., 1999; Mastepanov et al., 2013). As the net CH_4 flux is a result of production and oxidation, which both show a positive temperature response and a variable response in respect to WTL; it is not evident how the net CH_4 flux should respond to changing environmental conditions. At our site, the increasing temperature seemed to

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favor CH_4 uptake, and WTL had an additional control on the net uptake after the WTL had dropped below a certain threshold.

In addition to temperature, CH_4 uptake has been found to correlate with wind speed (Wang et al., 2013) and tree stand volume (Ojanen et al., 2010; Minkkinen et al., 2007). PAR has also been observed to correlate positively with CH_4 emissions in a *Sphagnum*-dominated mixed mire (Mikkelä et al., 1995). In this study, however, the correlations with PAR were low or absent, although u* correlated positively with the flux in some chambers in summer 2011 and also with the longer time periods when the flux data was grouped into u* classes of 0.05 m s^{-1}. This correlation with u* means that when the wind speed increased, the uptake of CH_4 decreased, which is in contrast to the findings of Wang et al. (2013) in Ontario, Canada. At our site, it is likely that there is CH_4 production deep in the soil (Pihlatie et al., 2010), but the CH_4 is oxidized by methanotrophic bacteria at the soil surface for most of the year and, additionally, are able to consume atmospheric CH_4. It is possible that with stronger winds the concentration gradient is mixed deeper in the soil, thus bringing more CH_4 from the deep soil layers to the atmosphere more rapidly, and the produced CH_4 is able to bypass oxidation by the bacteria.

5 Conclusions

This study presents two years of CH_4 exchange data measured by six soil chambers in a forestry-drained peatland. With the use of an accurate and precise gas analyzer, we were able to observe small CH_4 fluxes even during winter. Over the two year period, CH_4 exchange from the different chambers varied from –31.8 to –154.2 mg CH_4 m^{-2} yr^{-1} (average –67 mg CH_4 m^{-2} yr^{-1}). The site acted as a small annual CH_4 sink, although the effect of CH_4 emissions from the ditches was uncertain and it is possible that the site could be a small annual CH_4 source. There was clear seasonal variation in CH_4 flux and the site acted as a sink in summer. The sink decreased towards winter, and in early spring small periods of CH_4 emissions were recorded. In addition, diurnal variation was observed in summer, and the shape of the variation pattern varied between years. In summer 2011, the uptake was highest in the morning, but was highest in the afternoon in summer 2012. Net CH_4 uptake correlated best with soil temperature, especially at 20 and 30 cm depths, and also with WTL, showing increased CH_4 uptake with higher temperatures and lower WTL. Both linear and exponential regressions were used to calculate CH_4 fluxes, and like many studies before, we found that linear regression gave systematically lower flux estimates than the exponential regression. However, exponential regression was less useful for the estimation of low fluxes (sink/source < 3.5 µg CH_4 m^{-2} h^{-1}) due to the sensitivity of that regression method to signal-to-noise ratio. Therefore, we recommend using the flux calculated from linear regression on these low fluxes and the flux calculated from exponential regression when the flux is higher than the threshold. This threshold is dependent on the measurement system and cannot be generalized.

Data availability

The measured flux and meteorological data will be made available through European Fluxes Database Cluster (http://gaia.agraria.unitus.it/home).

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Table 1: Ground vegetation in each chamber and all-sided maximum vascular green area (VGA_{max}, m^2 vascular green surface m^{-2} forest floor).

Chamber	Vegetation	VGA_{max}
1	*Pleurozium schreberi* *Dicranum polysetum* *Vaccinium myrtillus*	2.04
2	*Pleurozium schreberi* *Dicranum polysetum* *Vaccinium vitis-idaea*	0.85
3	*Maianthemum bifolium* *Pleurozium schreberi* *Dicranum polysetum*	0.01
4	*Dryopteris carthusiana* Vaccinium myrtillus *Vaccinium vitis-idaea* *Pleurozium schreberi* *Dicranum polysetum*	2.34
5	*Pleurozium schreberi* *Dicranum polysetum*	0.11
6	*Sphagnum girgensohnii*	-

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Table 2: a) Seasonal (summer=JJA, autumn=SON, winter=DJF, spring=MAM) averages of CH_4 flux (µg CH_4 m^{-2} h^{-1}) calculated with linear and exponential regression with 95 % confidence intervals (±). The lowest row shows how much smaller the linear CH_4 flux was on average when compared to the exponential CH_4 flux. Values include the fluxes measured by all the six chambers. b) Same as in (a) but only includes fluxes with values > 3.5 µg CH_4 m^{-2} h^{-1}

a)

	Summer 2011	Summer 2012	Autumn 2011	Autumn 2012	Winter 11-12	Winter 12-13	Spring 2011	Spring 2012	4/11-3/13
Mean linear flux	-12.4 ±0.2	-9.5 ±0.3	-12.5 ±0.2	-9.1 ±0.2	-5.8 ±0.2	-0.9 ±0.1	-2.9 ±0.1	-3.1 ±0.2	-7.2 ±0.1
Mean exponential flux	-15.5 ±0.2	-12.2 ±0.4	-15.4 ±0.2	-11.3 ±0.2	-8.5 ±0.4	-2.6 ±0.6	-5.8 ±0.5	-5.4 ±0.5	-9.8 ±0.1
Number of closures	9114	2962	8376	8354	7228	7292	6338	4987	56579
Difference (%)	21.7 ±0.4	23.8 ±0.9	19.0 ±0.3	21.4 ±0.5	37.0 ±0.8	59.5 ±0.9	50.4 ±0.9	47.2 ±1.0	35.3 ±0.3

b)

	Summer 2011	Summer 2012	Autumn 2011	Autumn 2012	Winter 11-12	Winter 12-13	Spring 2011	Spring 2012	4/11-3/13
Mean linear flux	-13.1 ±0.2	-10.4 ±0.3	-12.7 ±0.2	-9.8 ±0.1	-8.4 ±0.2	-3.2 ±1.0	-6.0 ±0.2	-7.4 ±0.4	-10.8 ±0.1
Mean exponential flux	-16.1 ±0.2	-13.3 ±0.3	-15.6 ±0.2	-12.1 ±0.2	-11.9 ±0.4	-4.6 ±3.3	-9.1 ±0.5	-10.3 ±0.6	-13.7 ±0.1
Number of closures	8538	2592	8252	7360	4738	512	2518	1765	36195
Difference (%)	20.1 ±0.3	21.1 ±0.7	18.5 ±0.3	19.3 ±0.4	27.3 ±0.6	35.1 ±3.1	32.7 ±1.0	30.3 ±1.0	22.1 ±0.2

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Table 3: Average correlation coefficients (r) between hourly methane (CH_4) flux and environmental variables from six chambers during different seasons. **Bolded and underlined** r-values indicate that all six chambers had $p < 0.01$, **bolded** values indicate that five chambers had $p < 0.01$, underlined numbers indicate that 4 chambers had $p < 0.01$ and cursive numbers indicate that 3 chambers had $p < 0.01$. AirT=air temperature, ST=soil surface temperature, PAR=photosynthetically active radiation, u_*=friction velocity measured above the canopy, Tx=soil temperature at a depth of x cm, WTL=water table level. For clarity, negative r-values typically denote situations where CH_4 uptake increases when the value of explaining variable increases.

Season	AirT	ST	PAR	U*	T2	T5	T10	T20	T30	WTL*
Spring 2011	0.15	0.08	*-0.17*	-0.07	-0.22	**-0.33**	**-0.44**	**-0.53**	**-0.56**	**0.38**
Summer 2011	**0.04**	**0.12**	0.09	**0.25**	**0.17**	-0.16	**-0.27**	**-0.47**	**-0.55**	**0.49**
Autumn 2011	**-0.24**	**-0.17**	-0.06	**0.18**	**-0.14**	**-0.13**	-0.02	**-0.12**	**-0.14**	**0.31**
Winter 11-12	**-0.31**	**-0.20**	-0.09	-0.04	**-0.52**	**-0.56**	-0.11	**-0.63**	**-0.63**	**-0.63**
Spring 2012	**-0.43**	**-0.46**	-0.18	-0.04	**-0.46**	**-0.46**	*-0.07*	**-0.45**	**-0.46**	*0.04*
Summer 2012	**-0.50**	**-0.49**	-0.11	-0.09	**-0.52**	**-0.50**	0.04	**-0.46**	**-0.50**	**0.50**
Autumn 2012	**-0.54**	**-0.57**	0.11	0.10	**-0.58**	**-0.58**	**-0.14**	**-0.57**	**-0.56**	**0.60**
Winter 12-13	*0.05*	0.19	0.07	-0.05	-0.16	**-0.36**	**-0.42**	**-0.43**	**-0.43**	-0.11

*Negative WTL denotes water level below the soil surface, i.e. positive correlation results from increasing uptake with decreasing WTL.

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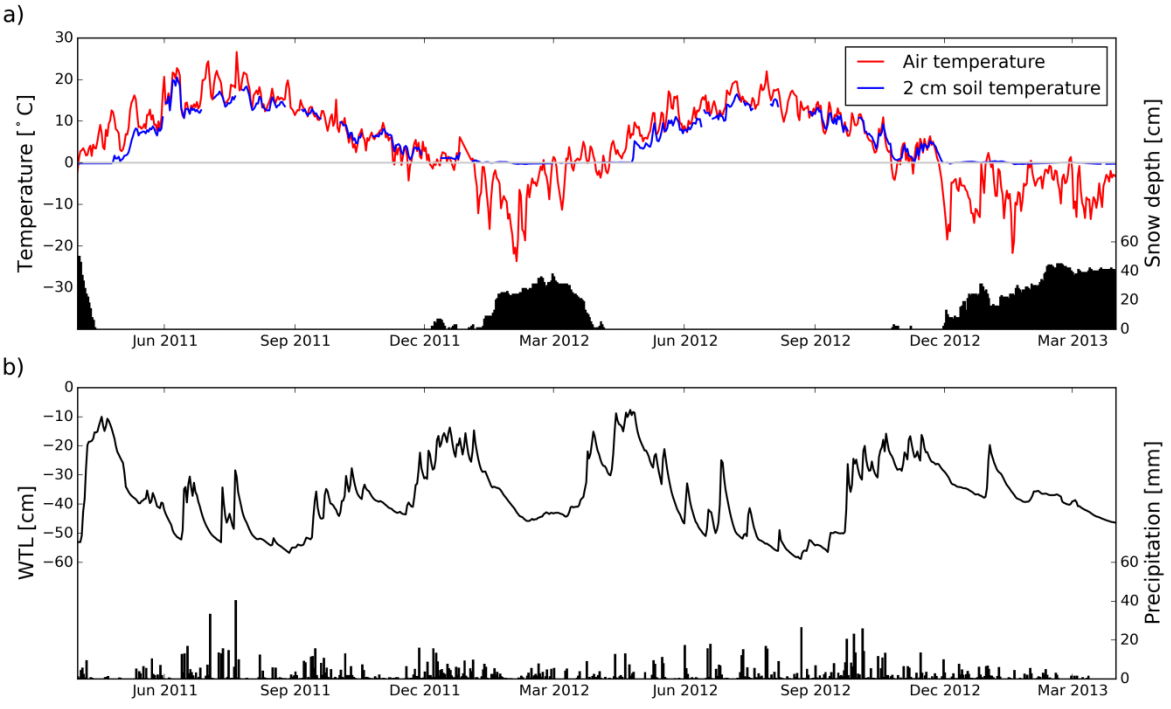

Figure 1: (a) The daily mean of air temperature (red) and soil temperature at 2 cm depth (blue) measured above the canopy at the site during the measurement period (April 2011 to end of March 2013), and the daily snow depth (bars) measured at the nearby Jokioinen observatory. (b) The daily mean water table (WTL) (line) from four different points at the site and the daily precipitation (bars) measured at the Jokioinen observatory.

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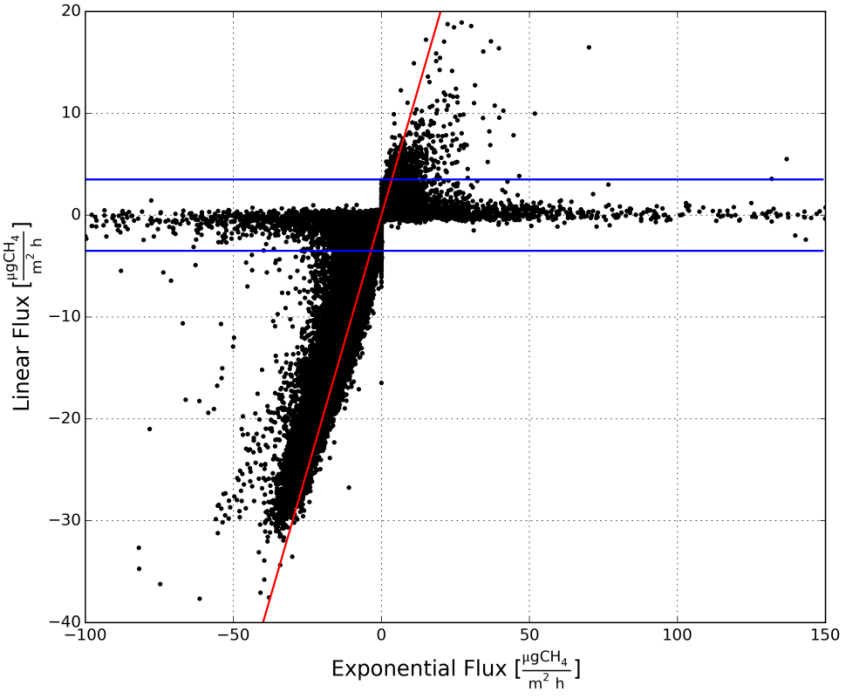

Figure 2: Flux based on linear regression (Linear Flux) versus flux based on exponential regression (Exponential Flux) over the two year measurement period. Red line describes the 1:1 relationship and blue lines are the ±3.5 µg CH_4 m^{-2} h^{-1} limits. If the datapoint was within the blue lines, linear regression was used instead of exponential regression.

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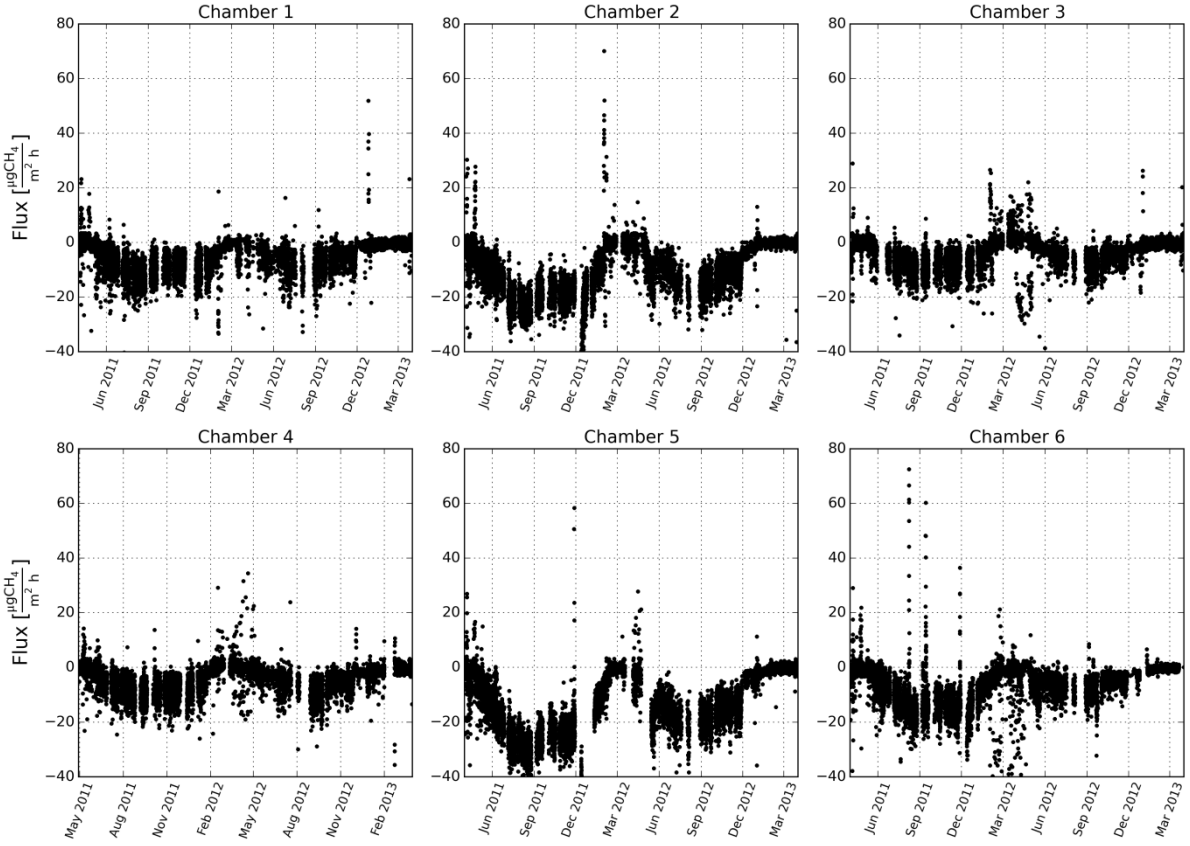

Figure 3: Hourly methane (CH_4) fluxes from April 2011 to March 2013 measured from each chamber. Negative values indicate uptake by the soil, positive values indicate emission to the atmosphere. Fluxes have been calculated using the exponential fit unless the value of the flux obtained from the linear fit was below 3.5 µg CH_4 m^{-2} h^{-1}.

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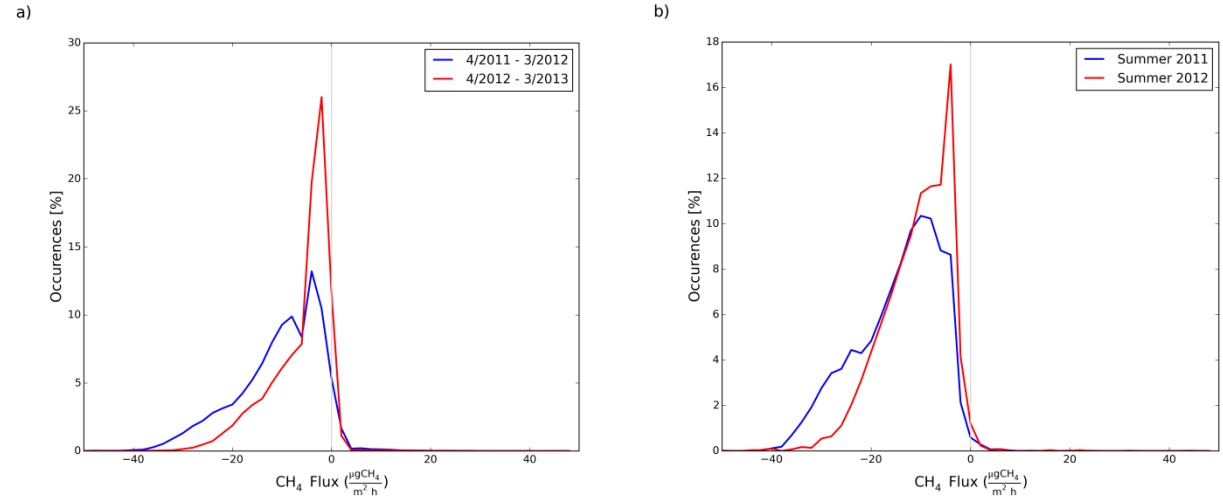

Figure 4: The flux rate distributions measured with all the chambers in (a) different years and (b) different summers. The flux rates were grouped into classes of 2 µg CH_4 m^{-2} h^{-1}.

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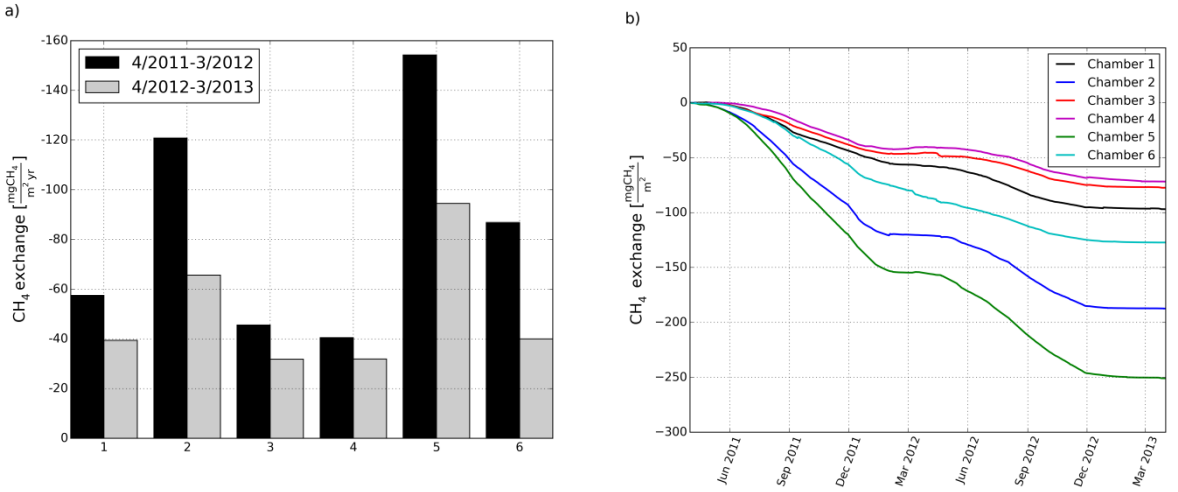

Figure 5: (a) Annual CH_4 exchange for each chamber for the two one-year monitoring periods, and (b) cumulative CH_4 exchange for each chamber starting from April 2011 until March 2013.

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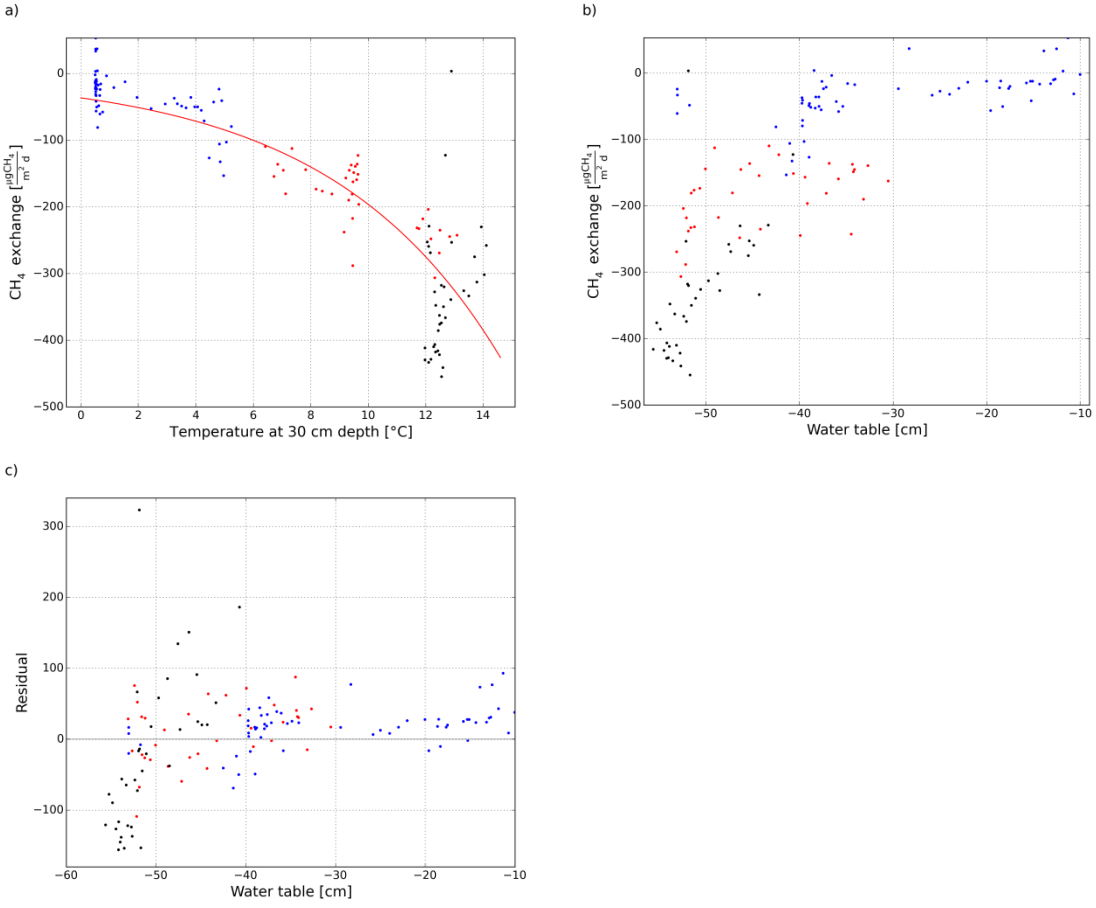

Figure 6: (a) Daily CH_4 exchange plotted against soil temperature at 30 cm in chamber #6 for spring (April and May) (blue), for the first half of the summer (June and part of July) (red) and for the second half of the summer (part of July and August) (black) in 2011. Red curve in (a) denotes the exponential fitting of Eq. 2 to the data. (b) Daily CH_4 exchange plotted against daily mean water table level (WTL) in chamber #6 using the same flux data as in panel (a). (c) Residuals of the exponential fitting (Eq. 2) in panel (a) plotted against WTL.

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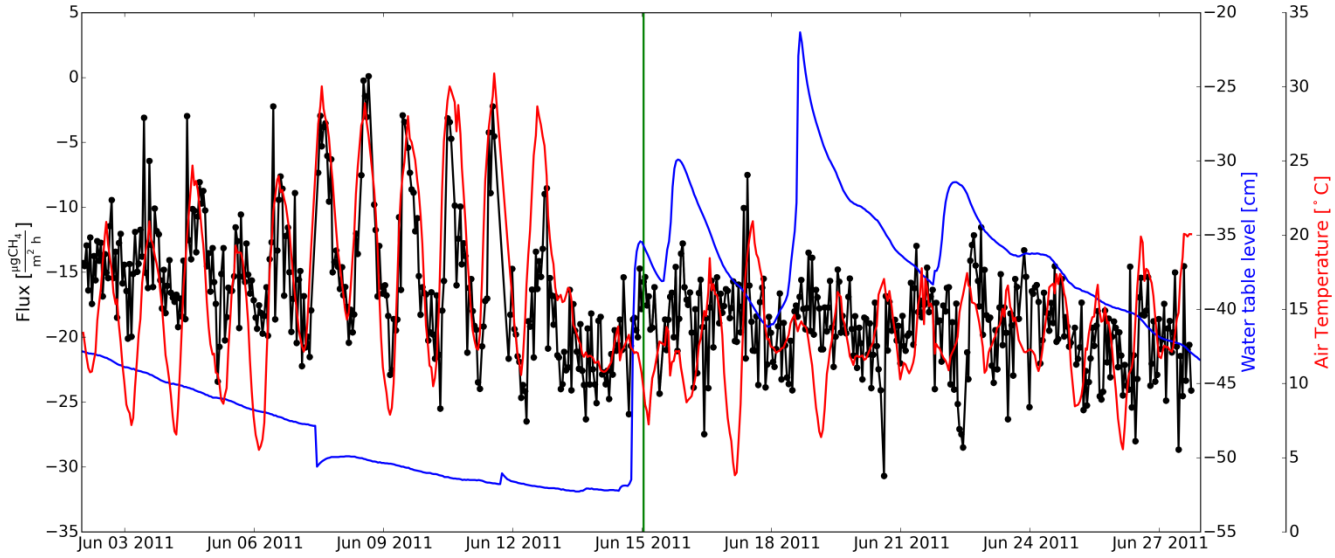

Figure 7: Hourly methane (CH_4) fluxes (black circles), air temperature (red curve) and water table level (WTL) (blue curve) in June 2011 measured by chamber #5. Green vertical bar shows where the data is split to panel groups of (a), (b), (c) and (d), (e), (f) in Fig 8.

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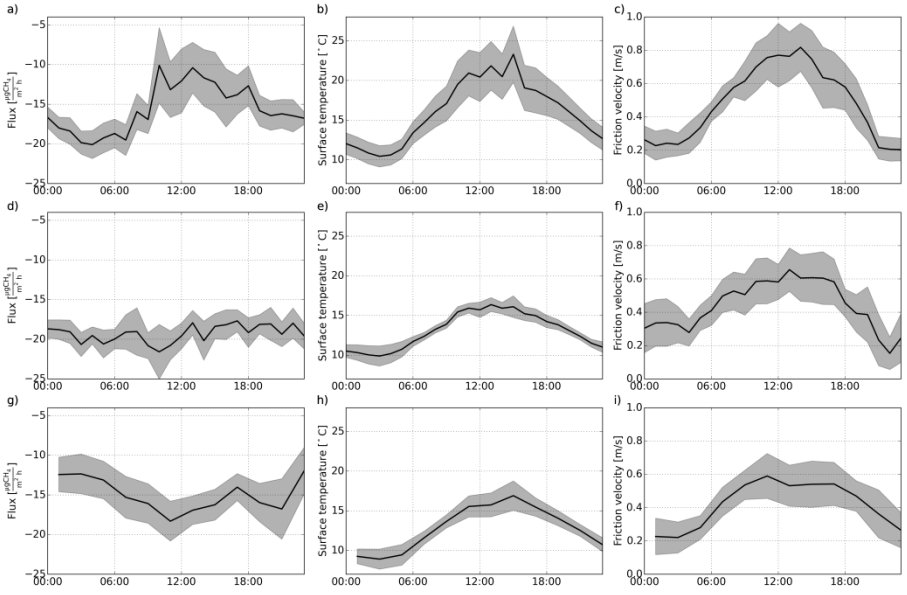

Figure 8: Diurnal variation of methane (CH_4) flux rate (a, d, g) and soil surface temperature (°C) (b, e, h) measured in chamber #5, and the friction velocity (u*) (c, f, i) measured above the canopy in 2-14 June 2011 (a, b, c), 15-27 June 2011 (d, e, f) and 15-30 June 2012 (g, h, i). Shading shows the 95 % confidence intervals.

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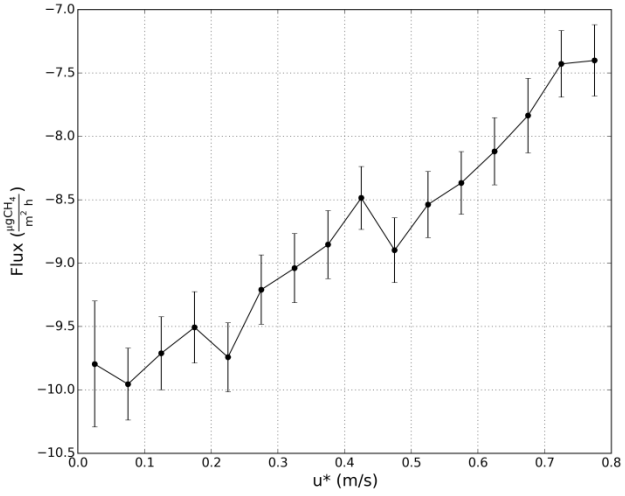

Figure 9. Average hourly methane (CH_4) flux over the whole measurement period against friction velocity (u*) measured above the canopy. Only negative fluxes (uptake by soil) are included. Data were grouped into u* classes of 0.05 m s^{-1} (n=1030-3658). Whiskers show 95 % confidence interval.