A snapshot of CO₂ and CH₄ evolution in a thermokarst pond near Igarka, northern Siberia

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Received 16 November 2007; revised 10 March 2008; accepted 14 April 2008; published 19 August 2008.

1. Introduction

[2] Thermokarst wetlands and ponds in the subarctic, which are located in land surface depressions resulting from permafrost melt, are strong sources of CH₄, but little is known about respiration processes supporting these emissions. We determined CH₄ fluxes and concentration profiles of dissolved gases and anions and some δ¹³C ratios of CO₂ and CH₄ in a thermokarst pond and adjacent smaller thermokarst depressions in the forest tundra near Igarka, northern Siberia in August 2006. Methane was emitted at 110–170 mg m⁻² d⁻¹ and produced mostly by CO₂ reduction, which also provided high Gibbs free energies on the order of 50–70 KJ mol⁻¹ H₂ due to high H₂ concentrations. The diffusive flux calculated from CH₄ gradients in the floating mat contributed <2% to emissions. CH₄ was apparently not oxidized deeper than 20 cm into the floating mat and the water body below. Anaerobic respiration required to reproduce nonsteady state CO₂ concentration maxima in the floating mat above the water body was 30–80 nmol cm⁻³ d⁻¹ or 250 mg m⁻² d⁻¹ and thus on a similar order of magnitude as CH₄ fluxes. The results suggest that floating mat-covered thermokarst ponds located in northern Siberian bogs effectively convert recently fixed carbon into CH₄ and thus allow for emissions independently from the finite, bog-derived carbon source. The relative contribution of recently fixed and old bog-derived carbon to C fluxes requires further investigation, however.


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0148-0227/08/2007JG000652S09.00
The study was conducted in the Little Grawijka Creek catchment located at 67°29.90′ latitude and 86°25.26′ longitude, about 10 km north of the city of Igarka, northern Krasnoyarsk Krai, Russian Federation. In nearby Igarka, the mean annual temperature (MAT) is −7.4°C and precipitation (MAP) 510 mm with half of it falling during July, August, and September [Lieth et al., 1999]. Snow cover usually extends into late May. Mean air temperature in August 2006 was 10.9°C. The catchment area is covered by loamy fluvioglacial sediments of the Karginskaya interstage (59 000 to 24 000 years B.P.) and limnic sediments of the Zyryansk epoch (122 000 to 59 000 years B.P.), on which raised bogs developed in the Holocene [Rodionov et al., 2007]. According to the geocryological map of Russia, the catchment area is located within the continuous permafrost belt [Ershov, 1996]. Our own catchment mapping revealed a rather discontinuous distribution of permafrost [Rodionov et al., 2007]. Considering also the relatively high permafrost temperature of −0.5 to −1.0°C (O.A. Kasansky, unpublished data), permafrost in this area is likely very sensitive to increases in MAT. In the raised bogs, permafrost degradation occurs in various forms ranging from small open pits about a meter deep and wide, over Sphagnum moss- and sedge-covered pools up to a few hundred m² surface area, to larger ponds with solidly developed floating mats (Figure 1). These small aquatic systems covered only 2.1% of the Little Grawijka Creek catchment but they determined the net-exchange of CH₄ of this landscape (Flessa et al., submitted manuscript, 2008).

2.2. Sampling and Analysis of Pore Water and Gases

[6] We selected a larger thermokarst pond (~500 m²) covered by a vegetation mat rich in Sphagnum spp., Eriophorum spp. and Scheuchzeria spp. for the analyses. The floating vegetation mat was connected with the sediment via long submersed shoots and roots, it was able to carry a person, was about 50 to 60 cm thick, and covered the entire water body. At this site ("Big Pond") we installed a boardwalk of 6 to 7 meter length prior to the measurements. Along the board walk, we installed 3 multilevel piezometers (MLP), at about 1m, 3m, and 6 m distance from the edge into the floating mat, the underlying water body, and a few centimeters into the underlying grayish sediment. The locations were chosen to cover a potential biogeochemical gradient from the bog surface into the pond. The water depth increased from about 130 cm (1m from the pond edge) to 200 cm (6 m from the edge). The submersed sediment consisted mainly of mineral material. Organic carbon content of the sediment sampled at a distance of 6 m from the edge of the pond was 113 g kg⁻¹ in 0 to 10 cm and 36 g kg⁻¹ in 10 to 30 cm (Flessa et al., submitted manuscript, 2008). The pond was completely covered by a compact ice layer from mid November to mid June. We do not have information on temperatures below the ice layer but extremely high CH₄ emissions after cracking of the ice indicate that the pond sediment was not frozen and CH₄ production continued during winter (Flessa et al., submitted manuscript, 2008). The pond was surrounded by steep banks consisting of bog material and it was connected with the Little Grawijka Creek by a small outflow. Deep cracks in the bog banks of the pond and erosion of thawed bog material into the pond indicate the ongoing process of thermokarst erosion. The MLPs consisted of plexiglas pipe segments, which were screwed together to provide a spatial sampling resolution of 10 cm to 20 cm down to depths of 140 cm (MLP “1 m”), 180 cm (MLP “6 m”), and 210 cm (MLP “3 m”) depth.

[7] The MLP segments were equipped with a crimp-vial (CS-Chromatographie, 13.7 mL volume) for pore-water sampling as described in Beer and Blodau [2007]. The crimp vials were filled with filtered drinking water and covered with a permeable polyamide membrane (Schleicher & Schuell, 0.45 μm pore size). The time for full equilibration between pore-water and interior solution is about
3 weeks but most of the equilibration is already completed after about 2 weeks time [Beer and Blodau, 2007], a period we resorted to due to time constraints of the field campaign. In addition, several other smaller water bodies differing in size and vegetation were equipped with MLPs to obtain information about dissolved CO2 and CH4, using wooden planks for access when necessary. These smaller water bodies encompassed a small open water pit (~6 m² surface), a partly Sphagnum moss covered pit (~15 m² surface), a fully Sphagnum moss covered pond (~30 m² surface), and a mostly Eriophorum spp. and Scheuchzeria spp. covered pond (~30 m² surface). The MLPs were inserted to maximum depth of 80–150 cm following predrilling with a sharpened birch trunk of similar diameter.

The MLP were sampled following retrieval of a few segments at a time to avoid degassing of the membrane filter stoppered vials. A volume of 8 mL of solution per crimp vial was carefully transferred with a syringe to empty crimp vials (13 mL volume), which were immediately butyl rubber stoppered. The syringe was equipped with a large diameter needle so that the solution could be quickly transferred without turbulence leading to degassing of the solution. After transfer to the University of Bayreuth, Germany, within about 3–4 days, previously dissolved CO2, CH4, and H2 were analyzed in the headspace of the vials. CO2 and CH4 concentrations were analyzed on a gas chromatograph, equipped with FID and a CO2 methanizer (8610C SRI Instruments, USA). Hydrogen (H2) partial pressures were analyzed on a H2-analyzer (Trace Analytical TA 3000r). As H2 partial pressures in the headspace ranged from 2 to 16 ppmV, the atmospheric background of H2 (0.55 ppmV) was uncritical. The original pore-water concentration was reconstructed using Henry’s law (KH CO2 = 3.89 10⁻² M atm⁻¹, KH CH4 = 1.3 10 M atm⁻¹, KH H2 = 7.8 10⁻⁴ M atm⁻¹) and accounting for both dissolved and gaseous gas concentrations in the vials. Measurements of 13C in CH4 and CO2 were conducted on a selected number of samples using a GC-MS (Combustion III interface, Supelco Carboxen-1010 PLOT column, Finnigan Delta-plus detector). Stable isotope compositions are reported in the standard δ notation: δ13C = ((Rsample/Rstandard) – 1) × 10³ with R = 13C/12C, and relative to V-DBP.

The remaining solution was filled into 10 mL PE-vials and frozen before analysis. Chloride and sulfate were analyzed by ion chromatography (Metrohm IC-System, Metrosep Anion Dual 2 (3) column at 0.5 (0.8) ml min⁻¹ flow rate, TCD with chemical suppression) and acetate by ion chromatography ( Dionex DX 500) as well. Limit of quantification (LOQ) for chloride and sulfate was ca. 5 μmol L⁻¹ and for acetate 10 μmol L⁻¹.

2.3. Calculations

A mean diffusive flux of CO2 and CH4 was calculated with Fick’s first law assuming a porosity of 0.95 and using diffusion coefficients taken from [Lerman, 1988], corrected for 8°C, and lumping together data from the top 30 cm of the measured concentration profiles of the three samplers at the main site. Net emissions of CH4 from the site “Big Pond” have been determined by Flessa et al., (submitted manuscript, 2008) who used closed chambers that were put on the vegetation mat. An apparent isotopic fractionation between CH4 and CO2 was calculated by α CO2,CH4 = (δ13CO2 + 1000) (δ13CH4 + 1000)⁻¹ and used to obtain information about microbial methanogenic pathways according to [Conrad, 2005; Whiticar et al., 1986]. The value of α CO2,CH4 represents the kinetic isotope fractionation effect for reduction of CO2 to CH4 as well as the influence of the shift in δ13CH4 due to the methane production from the acetoclastic and other pathways of methanogenesis. Gibbs free energy (ΔGf) (kJ mol⁻¹) of hydrogenotrophic and acetotrophic methanogenesis and homacetogenesis was estimated from the ratio of measured concentrations of participating substrates and products (equation 1), the standard Gibbs free energy of the reaction (ΔGf ⃗ (kJ mol⁻¹)), the gas constant R (8.31 10⁻³ kJ mol⁻¹ K⁻¹) and the absolute temperature T (K) using the Nernst equation (equation 1), as described in detail in [Beer and Blodau, 2007]:

$$\Delta G_f = \Delta G_f^\circ + R \cdot T \cdot \ln \frac{\prod_{i} (\text{products})^{vi}}{\prod_{i} (\text{substrates})^{vi}}$$

Pressure dependencies and effects of ionic strength were ignored. A concentration of 5 μmol L⁻¹ (half the limit of quantification, LOQ) was used for thermodynamic calculations when acetate was not detectable, which was the case for most samples.

Pore water profiles of CO2 and CH4 in the floating mat of the Big Pond site were not in steady state (see results) and we thus could not use inverse pore water modeling to infer respiration rates, as was done before elsewhere [Beer and Blodau, 2007; Blodau et al., 2007a]. To estimate a range of CO2 production necessary to reproduce the observed concentration depth profiles we converted a multiple box model, implemented in Stella©

Figure 1. Thermokarst features in the sampled bog Little Grawikja Creek catchment. (a) An initial pitlike depression of about 80-cm depth (“Pit Pond”). (b) A sedge covered pond (“Sedge pond”) during installation of a multilevel piezometer (MLP) from a wooden plank. (c) Sphagnum dominated small pond (“Dark pond”). (d) Main site (“Big Pond”) with boardwalk and R. Rees.
system dynamics software, describing vertical diffusion and advection of chloride [Beer and Blodau, 2007], for CO2 production and transport. The model follows the law of mass conservation (equation 2) with \( F_z \) the diffusive flux (\( \mu \text{mol cm}^{-2} \text{d}^{-1} \)) in direction \( z \), \( C \) the concentration of dissolved CO2 (\( \mu \text{mol cm}^{-3} \)), \( n \) the effective porosity, and \( R \) the production of CO2 (\( \mu \text{mol cm}^{-3} \text{d}^{-1} \)).

\[
\frac{\partial F_z}{\partial z} = -n \frac{\partial C}{\partial t} + R
\]  

(2)

\( R \) was adjusted until simulated and measured concentration data matched, see Results and Discussion section. Diffusion was calculated using Fick’s first law and advection was set to 0.

3. Results and Discussion

3.1. Methane

[12] We have reported that the “Big Pond” site was a strong source of atmospheric CH4 with net emissions of CH4 averaging 20 g \( \text{m}^{-2} \text{yr}^{-1} \) and active emissions occurring from mid June to early November (Flessa et al., submitted manuscript, 2008). These measurements were carried out with the static chamber technique and included both diffusive and plant mediated transport. Emissions were calculated from a linear concentration increase with \( R^2 > 0.85 \) and thus also included ebullition that did not lead to violation of this criterion. Emissions started with the disintegration of the ice layer during spring thaw and ended after the formation of a compact ice layer in November. Emissions at the time of pore water sampling in August 2006 were 110–170 mg CH4 \( \text{m}^{-2} \text{d}^{-1} \), which is close to average daily emissions from wet ombrotrophic sites at the Stordalen mire in northern Sweden [Christensen et al., 2004], thermokarst wetlands at Bonanza Creek, Alaska [Wickland et al., 2006] and in Manitoba, Canada [Bubier et al., 1995], and thaw lakes in North East Siberia [Walter et al., 2006]. Emissions were thus similar to other thermokarst wetlands and lakes and high compared to CH4 emissions across peatland ecosystems, exceeding median emission up to an order of magnitude or more [Blodau, 2002]. In the Little Grawijka catchment, surrounding bogs that are underlain by permafrost and are “dry” during summer even acted as a weak CH4 sink (Flessa et al., submitted manuscript, 2008).

[13] In agreement with the large fluxes, we recorded maximum CH4 concentrations of 0.53 to 0.97 mmol L\(^{-1}\) at depths of 20 to 40 cm (Figure 2) at the Big Pond site and similar CH4 concentrations also in other thermokarst depressions (Figure 3). At the “Big Pond” site, CH4 was likely transported by aerenchymatic roots of Eriophorum and other emergent vegetation, which act as effective conduits for CH4 [Frenzel and Rudolph, 1998; Shannon et al., 1996]. Ebullition may have further contributed to emissions as the CH4 partial pressures in the pond exceeded 0.2 atm, equivalent to ca. 0.39 mmol L\(^{-1}\) at 8°C, which is initially sufficient for ebullition to occur [Fechner-Levy and Hemond, 1996]. It has been previously documented that accelerated transport by root conduit transport can lead also to a lowering of CH4 pools in the wetland soils. In the study by Shannon et al. [1996] emissions of 200 to 400 mg C \( \text{m}^{-2} \text{d}^{-2} \) were reached in a temperate wetland with a dense cover of Scheuchzeria palustris and fairly uniform CH4 concentrations ranging from 300 to 500 \( \mu \text{mol L}^{-1} \), which is lower than in the floating mat at the “Big Pond” site. In the study by Shannon et al. [1996] and studies cited therein, 60 to 90% of the CH4 was emitted through aerenchymatic transport and some additional contribution of ebullition, whereas diffusion was subordinate. In agreement with such findings the estimated diffusive fluxes at the “Big Pond” site only reached 3.5 mg m\(^{-2} \) d\(^{-1}\) or less than 2% of the total CH4 emission.

[14] Information about the predominating pathways of CH4 production can be gained from both the isotopic composition of produced CH4, as well as thermodynamic

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**Figure 2.** Dissolved concentrations of CO2, CH4, H2, sulphate, and chloride in MLPs located approximately 1 m, 3 m, and 6 m from the edge of the “big Pond” site.

- Big Pond-1m
- Δ · · · Big Pond-3m
- Big Pond-6m
considerations [Conrad, 1999, 2005; Whiticar, 1999]. CH₄ produced by acetate cleavage is usually not as depleted in ¹³C as CH₄ produced from CO₂ reduction with H₂, resulting in apparent fractionation factors of <1.055 versus >1.065 [Conrad, 2005; Whiticar et al., 1986]. Even in experiments with cultured microorganisms, fractionation factors are quite variable and influenced by environmental conditions, such as temperature, substrate concentrations, and Gibbs free energies available for the process [Conrad, 2005]. A higher fractionation factor for hydrogenotrophic methanogenesis in the range of 1.07 to 1.09 can for example be expected when Gibbs free energies for the process are <40 KJ mol⁻¹ CH₄ [Penning et al., 2005]. It has been also documented that diffusive transport in aerenchymatic plants leads to preferential emission of ¹³C. At a field site in Florida this process lead to an increase in δ¹³C of 10.6 ± 3.7‰ in emitted CH₄ compared to pore water sedimentary CH₄ [Chanton and Whiting, 1996] which would entail an increase in the derived apparent fractionation factor of methanogenesis. Without further information, the diagnostic power of the apparent fractionation factor is thus limited and can only be used as a coarse indication of the contribution of methanogenic pathways to CH₄ production [Conrad, 2005]. In both fens and bogs, CH₄ has been found to be increasingly depleted in ¹³C deeper into the peat [Hornibrook et al., 1997; Popp et al., 1999]. Accordingly it has been suggested that in the biologically active uppermost layers, where fresh substrates are available and fluctuations in temperature and redox conditions are frequent, acetotrophic methanogenesis is generally more important and gives way to hydrogenotrophic methanogenesis at greater depths [Hornibrook et al., 1997].

[15] We determined δ¹³C values of −11.8 to −25.7‰ for CO₂ and −56.2 to −62.4‰ for CH₄ (Figure 4) resulting in apparent fractionation factors of 1.039 to 1.060. Such intermediate values suggest that both acetate cleavage and CO₂ reduction contributed to CH₄ production at the “Big Pond” and two other sites, which were characterized by fractionation factors of 1.039 to 1.051. In agreement with this interpretation, acetate, which is a ubiquitous intermediate of organic matter degradation [Conrad, 1999], did not accumulate in the pore water and was mostly below LOQ. Conversion of acetate into CH₄ thus did not appear to be inhibited as for example reported by Duddleston et al. [2002] for a bog in Alaska. Particularly at the “Big Pond” site, the CO₂ was also enriched in ¹³C compared to organic matter, which is generally characterized by a ¹³C of about −25 to −30 ‰ (median ~27‰) when produced by C3 plants [O’Leary, 1988]. This finding supports a reduction of CO₂, since during this process the remaining pool of CO₂ is left enriched with the heavier isotope. In contrast to previous findings [Hornibrook et al., 1997; Popp et al., 1999] we could not find evidence for a substantial shift in methanogenic pathways from the floating mat, consisting mostly of recently fixed organic matter, into the underlying water body, containing an unknown mixture of “old” and “new” carbon. In the investigated ponds CO₂ reduction was apparently more important than in incubated organic matter samples from “dry” bog locations. In a previous study, acetotrophic methanogenesis dominated at summer soil temperatures when bog samples taken from the catchment were incubated after a storage period of 4 months [Metje and Frenzel, 2007]. The exact reason why acetotrophic methanogenesis did not predominate more clearly in the floating mat, despite the availability of recently fixed organic matter, cannot be identified. One may speculate that permanently water saturated conditions and high partial pressures of CO₂ and H₂, which favor CO₂ reduction thermodynamically, are factors that encouraged hydrogenotrophic methanogenesis in the ponds compared to other wetland environments.

[16] Both pathways were thermodynamically viable processes, as Figure 5 illustrates. Hydrogen concentrations reached between 50 and 900 nmol L⁻¹ (Figure 2) and resulted in a quite large driving force for hydrogenotrophic methanogenesis on the order of 50 to 70 KJ mol⁻¹ (substrate). The driving force for acetotrophic methanogenesis
was considerably smaller on the order of 26 to 38 kJ mol\(^{-1}\) (acetate) but still larger than threshold values determined in slurry incubation experiments with rice paddy soil [Chin and Conrad, 1995], and assuming that acetate occurred at half LOQ of 10 \(\mu\)mol L\(^{-1}\). In contrast, the energy gain from homoacetogenesis of CO\(_2\) and H\(_2\) to acetate was closer to 0 with values of 7 to 17 kJ mol\(^{-1}\). The fact that hydrogen concentrations were high suggests that 1) hydrogenotrophic methanogenesis was not able to lower hydrogen concentrations to thermodynamically controlled steady state levels and 2) that alternative electron acceptors were not able to compete effectively for hydrogen. The reasons for the high H\(_2\) levels cannot be clarified. In absence of alternative electron acceptors hydrogenotrophic methanogenesis often operates at lower H\(_2\) concentration levels, resulting in a \(\Delta G\) of \(-20\) to \(-40\) kJ mol\(^{-1}\) (CH\(_4\)) in experiments with freshwater soils and sediments [Chin and Conrad, 1995; Conrad, 1999; Rothfuss and Conrad, 1993]. In their presence and with rapid utilization, hydrogen concentrations are typically lowered to steady state values of \(<0.1\) to \(2\) \(\mu\)mol L\(^{-1}\) thus resulting in a positive \(\Delta G\) of hydrogenotrophic methanogenesis [Heimann et al., 2007; Lovley and Goodwin, 1988].

[17] In agreement with the high H\(_2\) concentrations and a subordinate importance of alternative terminal electron accepting processes, nitrate was generally not detectable and sulfate concentrations in the Big Pond site were \(<40\) \(\mu\)mol L\(^{-1}\) in the surface layer of all three wells (Figure 2). A concentration minimum of \(<10\) \(\mu\)mol L\(^{-1}\) was reached at a depth of 30 to 40 cm. This suggests that some sulfate was reduced in the floating mat itself and either provided from the surface (“1 m” and “3 m”), e.g., from precipitation and reoxidation of reduced sulfur [Blodau et al., 2007b], and additionally from the water body and deeper layers below (“6 m”). The latter is also confirmed by the chloride profiles which indicated a deeper source of more mineralized water particularly at the “6 m” site (Figure 2).

[18] The fairly constant \(\delta^{13}C\) values of CH\(_4\) in the profiles of the ponds (Figure 4) also suggests that CH\(_4\) was not oxidized to a significant degree by aerenchymatic oxygen transport in the rhizosphere, which has been described and analyzed elsewhere [Arah and Stephen, 1998]. Methane oxidation is associated with increasing \(\delta^{13}C\) values due to enrichment of the heavier isotope in the remaining CH\(_4\) pool, which generally leads to increasing \(\delta^{13}C\) CH\(_4\) values where oxygen is available [Whiticar, 1999], e.g., near the water table. We did not obtain \(\delta^{13}C\) values near the surface of the floating mat but up to 20 cm below surface \(\delta^{13}C\) values did not increase. Significant CH\(_4\) oxidation was thus likely restricted to the aerated organic uppermost layers, possibly Sphagnum mosses [Basiliko et al., 2004] and root conduits, although in Eriophorum oxidation was shown to be very slow [Frenzel and Rudolph, 1998].

### 3.2. Carbon Dioxide

[19] Concentrations of CO\(_2\) were about an order of magnitude higher than concentrations of CH\(_4\) and reached a maximum of 6.9 mmol L\(^{-1}\) in Big Pond and between 5 and 12 mmol L\(^{-1}\) in the other sampled thermokarst depressions (Figures 2 and 3). Similar CO\(_2\) concentrations have been reported from thermokarst wetlands at the Bonanza Creek field site using pore water peepers [Wickland et al., 2006]. The resulting diffusive flux of CO\(_2\) determined by Fick’s law and the pore water profiles in “Big Pond” was 20.5 mg C m\(^{-2}\) d\(^{-1}\) and thus only about 15 % of the chamber CH\(_4\) efflux in August 2006. This finding raises the question whether anaerobic decomposition in the floating mat was capable of producing the emitted CH\(_4\), given that decomposition of organic matter with an oxidation state of 0, such as polysaccharides, requires the release of equivalent amounts of CO\(_2\). To gain further insight into this question we first examined the shape of the CO\(_2\) concentration profiles. Both CH\(_4\) and CO\(_2\) concentrations peaked in “Big Pond” wells “2 m” and “3 m” in the floating mat and at intermediate depths also at the other sites. This first of all indicates a stronger respiration activity that was likely caused by higher “soil” temperatures near the surface, the input of labile organic matter from roots and possibly root respiration. Secondly, such concentration profiles are also an indication of nonsteady state between production and concentrations, as in the long run production coupled to a zero-flow boundary at the bottom must lead to monotonously increasing concentrations with depth [Berner, 1980]. The accelerated production during summer did apparently not equilibrate with the pool size present in the floating mat and underlying water and the calculated diffusive flux underestimated the true CO\(_2\) production. To infer the magnitude of nonsteady state CO\(_2\) production in the floating mat we thus simulated the CO\(_2\) depth profiles.

[20] To estimate CO\(_2\) production required to generate the CO\(_2\) maximum in the floating mat we ran two simulations, 1) to create the profile in the lower part of the pond (<60 cm depth) assuming an arbitrary accumulation period of 10 years and 2) based on this profile, to create the concentration maximum in the floating mat within a three month period (May to August). The rationale behind this approach was the assumption that absence of autotrophic respiration and slowness of heterotrophic production in the cold season as well as gradual freezing of the floating mat (<60 cm depth) would result in low CO\(_2\) concentrations in spring so that the pool in the floating mat had to build up

![Figure 5. Gibbs free energies (\(\Delta G\)) of hydrogenotrophic and acetotrophic methanogenesis and homoacetogenesis calculated from gaseous and dissolved concentrations in MLPs installed in “Big Pond.”](image-url)
agreement with this finding, the simulated nonsteady state anaerobic respiration rates in the water saturated floating mat were fairly small compared to rates of net ecosystem production on the order of a few hundred g C m\(^{-2}\) yr\(^{-1}\) that are characteristic of northern wetlands [Bubier et al., 1998]. As has been pointed out, even small relative changes in the area covered by thermokarst wetlands, ponds, and lakes may dramatically raise CH\(_4\) emissions on the landscape scale [Christensen et al., 2004; Flessa et al., submitted manuscript, 2008; Wickland et al., 2006]. The consequences for CH\(_4\) emissions cannot be quantified yet due to a lack of more detailed information on the processes supporting carbon cycling in these systems, their future areal extension, their life cycle, and cumulative CH\(_4\) and CO\(_2\) balance over the life cycle.

[22] Acknowledgments. The support of the Deutsche Forschungsgemeinschaft (DFG) and of the German Ministry of Science and Education (BMBF) to H. Flessa, G. Guggenberger, and C. Blodau is gratefully acknowledged. We thank Martina Heider for laboratory assistance, Pjotr Karas and Alexander Tiunov for all their help with field work, and Swetlana Poljuhova (Field Station Igarka of the Permafrost Institute Yakutsk) for excellent laboratory analyses and help with respect to logistics.

References


Figure 6. Model fits of the mixed reservoir model used to estimate nonsteady state anaerobic CO\(_2\) production in the floating mat during summer. “Scenario 1” was run using a production of 1 nmol cm\(^{-3}\) d\(^{-1}\) in the deeper layer for 10 years. “Scenario 2” was run for a time period of three months with production rates of 30 to 80 nmol cm\(^{-3}\) d\(^{-1}\) in the floating mat using the “Scenario 1” concentration profile as starting condition.

with the beginning of summer. A reasonable fit was obtained using a long-term production rate of 1 nmol cm\(^{-3}\) in the water body below and a short-term production rate of 30 to 80 nmol cm\(^{-3}\) d\(^{-1}\) in the floating mat (Figure 6). The resulting cumulative 3 month production of CO\(_2\) in the floating mat was 23 g C m\(^{-2}\), which was on the order of annual CH\(_4\) emissions from the surface. Such an ad hoc simulation of a single date of concentration profiles is of course subject to considerable uncertainty; but the order of magnitude of respiration rates may be adequately reflected. If so, nonsteady state anaerobic respiration in the recently formed floating mat was capable of sustaining the determined CH\(_4\) fluxes from the pond.

4. Conclusions

[21] The depth distribution of the CO\(_2\) and CH\(_4\) concentrations and our simulation exercise suggest that the floating mat strongly contributed to summer anaerobic respiration and methanogenesis in a northern Siberian thaw pond. Methane was probably mostly produced by the hydrogeno-trophic and subordinately by the acetotrophic pathway. Both processes provided sufficient free energy and isotopic fractionation factors were in a range suggesting a greater importance of CO\(_2\) reduction. Methane was apparently not oxidized deeper in the floating mat according to the \(^{13}\)C isotopic data. Previously it has been reported that the formation of wet thermokarst wetlands and ponds results not only in increasing CH\(_4\) emissions but also in an increasing CO\(_2\) sink on the landscape scale [Camill et al., 2001; Christensen et al., 2004; Wickland et al., 2006]. In


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