Partitioning carbon fluxes in a permafrost landscape

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Tim Eckhardt
aus
Hamburg

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# List of Abbreviations and Symbols

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<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha$</td>
<td>initial canopy quantum efficiency</td>
</tr>
<tr>
<td>$\text{AIC}_c$</td>
<td>Akaike Information Criterion</td>
</tr>
<tr>
<td>$\text{ALD}$</td>
<td>active layer depth</td>
</tr>
<tr>
<td>a.s.l.</td>
<td>above sea level</td>
</tr>
<tr>
<td>$C$</td>
<td>carbon</td>
</tr>
<tr>
<td>$\text{CH}_4$</td>
<td>methane</td>
</tr>
<tr>
<td>$\text{CH}_3\text{COOH}$</td>
<td>acetic acid</td>
</tr>
<tr>
<td>$\text{CO}_2$</td>
<td>carbon dioxide</td>
</tr>
<tr>
<td>$\text{CO}_2$-eq</td>
<td>carbon dioxide equivalent</td>
</tr>
<tr>
<td>$C/N$</td>
<td>carbon to nitrogen ratio</td>
</tr>
<tr>
<td>$\text{EC}$</td>
<td>Eddy covariance</td>
</tr>
<tr>
<td>$\text{GPP}$</td>
<td>gross primary productivity</td>
</tr>
<tr>
<td>$\text{GWP}$</td>
<td>global warming potential</td>
</tr>
<tr>
<td>$H_2$</td>
<td>hydrogen</td>
</tr>
<tr>
<td>$H_2O$</td>
<td>water</td>
</tr>
<tr>
<td>$Hz$</td>
<td>hertz</td>
</tr>
<tr>
<td>LAI</td>
<td>leaf area index</td>
</tr>
<tr>
<td>mg</td>
<td>milligram ($= 10^{-3}$ g)</td>
</tr>
<tr>
<td>n</td>
<td>number of replicates</td>
</tr>
<tr>
<td>N</td>
<td>nitrogen</td>
</tr>
<tr>
<td>NDVI</td>
<td>normalized difference vegetation index</td>
</tr>
<tr>
<td>$\text{NEE}$</td>
<td>net ecosystem exchange</td>
</tr>
<tr>
<td>$\text{NPP}$</td>
<td>net primary productivity</td>
</tr>
<tr>
<td>$p$</td>
<td>probability value</td>
</tr>
<tr>
<td>$\text{PAR}$</td>
<td>photosynthetic active radiation</td>
</tr>
<tr>
<td>$Pg$</td>
<td>petagram ($= 10^{15}$ g)</td>
</tr>
<tr>
<td>pH</td>
<td>negative decadal logarithm of hydrogen ion activity</td>
</tr>
<tr>
<td>$P_{max}$</td>
<td>maximum canopy photosynthetic potential</td>
</tr>
<tr>
<td>ppm</td>
<td>parts per million</td>
</tr>
<tr>
<td>$Q_{10}$</td>
<td>ecosystem sensitivity to changing temperatures</td>
</tr>
<tr>
<td>Symbol</td>
<td>Definition</td>
</tr>
<tr>
<td>--------</td>
<td>------------</td>
</tr>
<tr>
<td>$R^2$</td>
<td>coefficient of determination</td>
</tr>
<tr>
<td>$R^2_{adj}$</td>
<td>adjusted coefficient of determination</td>
</tr>
<tr>
<td>$R_A$</td>
<td>autotrophic respiration</td>
</tr>
<tr>
<td>$R_{base}$</td>
<td>basal respiration</td>
</tr>
<tr>
<td>$R_{eco}$</td>
<td>ecosystem respiration</td>
</tr>
<tr>
<td>$R_H$</td>
<td>heterotrophic respiration</td>
</tr>
<tr>
<td>RMSE</td>
<td>root mean square error</td>
</tr>
<tr>
<td>RVI</td>
<td>ratio vegetation index</td>
</tr>
<tr>
<td>SOC</td>
<td>soil organic carbon</td>
</tr>
<tr>
<td>SOM</td>
<td>soil organic matter</td>
</tr>
<tr>
<td>SMR</td>
<td>soil microbial respiration</td>
</tr>
<tr>
<td>$T_{air}$</td>
<td>air temperature</td>
</tr>
<tr>
<td>$T_g$</td>
<td>teragram ($= 10^{12}$ g)</td>
</tr>
<tr>
<td>TIC</td>
<td>total inorganic carbon</td>
</tr>
<tr>
<td>$T_{soil}$</td>
<td>soil temperature</td>
</tr>
<tr>
<td>$T_{surf}$</td>
<td>surface temperature</td>
</tr>
<tr>
<td>TOC</td>
<td>total organic carbon</td>
</tr>
<tr>
<td>VWC</td>
<td>volumetric water content</td>
</tr>
<tr>
<td>WT</td>
<td>water table</td>
</tr>
<tr>
<td>$\mu g$</td>
<td>microgram ($= 10^{-6}$ g)</td>
</tr>
</tbody>
</table>
Summary

Arctic tundra ecosystems have acted as a historical sink for atmospheric carbon (C). Therefore, these ecosystems contain significant amounts of carbon in their frozen soils and sediments. The function as strong C sink is a product of short summers, low temperatures, well-adapted vegetation and water-saturated conditions, which causes low degradation and mineralization rates of soil organic matter (SOM). The Arctic is currently facing amplified climate warming and the rate of this warming is expected to accelerate. A number of changes are predicted in response to warming. On the one hand, warming is expected to cause the deeper thaw of permafrost-affected soils leading to both the enhanced production of carbon dioxide (CO$_2$) due to the increasing degradation of SOM, and increased methane (CH$_4$) formation, which could create a positive climate feedback to global warming. On the other hand, due to higher temperatures the assimilation of CO$_2$ by vegetation will increase. Therefore, it remains uncertain how the C sink function of arctic tundra landscapes will react to future changes in climate.

This study aimed to determine the CO$_2$ and CH$_4$ fluxes at the soil-plant-atmosphere interface in an arctic tundra ecosystem and to identify the main environmental drivers of these fluxes. Furthermore, as the processes governing CO$_2$ net ecosystem exchange (NEE) react differently on a changing climate, the CO$_2$ fluxes were partitioned into gross primary productivity (GPP) and ecosystem respiration ($R_{eco}$). $R_{eco}$ was further partitioned into its autotrophic ($R_A$) and heterotrophic respiration ($R_H$) components. The study was conducted using chamber measurements on the microscale (1 m – 10 m) in a polygonal tundra environment in the Lena River Delta in 2014 and 2015. In order to estimate the CO$_2$ fluxes over the complete growing season in 2015, they were reproduced using flux models.

This work finds the polygonal tundra in the Lena River Delta to be a robust sink for atmospheric CO$_2$ and a source for CH$_4$. The CO$_2$ sink strength was highly heterogeneous on the microscale. The net CO$_2$ uptake at a wet-depressed polygonal center over the 2015 growing season was more than twice as high as measured at a drier polygonal rim. In addition to higher GPP fluxes, the differences in NEE between the two microsites were primarily caused by lower $R_{eco}$ fluxes at the center compared to the rim. Here, the contrasting hydrological conditions cause the CO$_2$ flux differences between the microsites,
where high water levels lead to lower decomposition rates due to anoxic conditions. For the first time, the different response of \textit{in situ} measured \( R_A \) and \( R_H \) fluxes to hydrological conditions was determined. It was shown that a high water table can lower \( R_A \) fluxes, most likely due to the submersion of plants; while a low water table lead to enhanced \( R_H \) fluxes and can desiccate mosses, therefore lowering \( R_A \) fluxes. The \( R_H \) fluxes presented in this study are the first \textit{in situ} measured \( R_H \) fluxes, which cover an almost complete growing season from the vast Russian arctic tundra.

The environmental parameter with the highest explanatory power for the CH\(_4\) fluxes was the active layer depth (ALD), although in general CH\(_4\) fluxes were poorly correlated with environmental parameters, which highlights the complicated identification of factors influencing CH\(_4\) fluxes. In contrast to previous studies, the dry tundra (polygonal rim) was not found to act as sink for atmospheric CH\(_4\), most probably due to the low soil temperatures and high substrate availability at the study site.

The calculation of CO\(_2\) and CH\(_4\) budgets for the polygonal tundra revealed that the polygonal rims were a stronger net CO\(_2\) sink than the centers in 2015 due to the high coverage of dry tundra at the study site. Considering the global warming potential (GWP) of CH\(_4\) compared to CO\(_2\), the radiative forcing of CH\(_4\) emissions were five-fold lower in magnitude than the strength of the CO\(_2\) sink in 2015. However, in 2014 both the CO\(_2\) and the CH\(_4\) budgets were marginally positive, denoting that the area functioned as a net source for atmospheric C. This finding highlights the pronounced inter-annual variability of C fluxes between the soils and atmosphere at the study site.

The partitioning of CO\(_2\) fluxes provides unique insights into the individual contributions of each of the processes governing the CO\(_2\) NEE. Improved knowledge of these contributions is of crucial importance to gain a better understanding of the reaction of arctic ecosystems to changing climatic conditions as these processes react differently to environmental parameters. Estimates of the CH\(_4\) source strength of different microsites in the arctic tundra improves estimates of the C budgets from these contrasting environments. Furthermore, the presented results provide an improved understanding of the response of CO\(_2\) and CH\(_4\) fluxes to environmental controls from different arctic tundra sites, which is needed to optimize model simulations of future C fluxes.
Zusammenfassung


Die polygonale Tundra stellt eine robuste CO₂-Senke und eine CH₄-Quelle dar. Beide Austauschflüsse zeigen eine hohe Heterogenität auf kleinräumlicher Skala (1 – 10 m): In einem wassergesättigten Polygonzentrum wurde eine mehr als doppelt so hohe CO₂- Aufnahme ermittelt als auf einem verhältnismäßig trockenen Polygonwall. Neben der höheren Primärproduktion führen hauptsächlich Unterschiede in der


1. Introduction and objectives

The amount of carbon (C) stored in northern high-latitude permafrost soils is with more than 1300 Petagram (Pg) distinctly higher as the current amount of C in the atmosphere (832 Pg) (Ciais et al., 2013, Hugelius et al., 2014). Of this 1300 Pg, about 500 Pg are stored in the seasonally thawed active layer, while ~800 Pg are stored in perennially frozen ground (Hugelius et al., 2014). Given this large amount of C stored in permafrost soils, the response of the Arctic C cycle to a changing climate is of global importance (McGuire et al., 2009). Over thousands of years the C was stored in permafrost soils due to a long freezing period and poor drainage, resulting in water-saturation (Ping et al., 2015). Since a few decades, arctic ecosystems are facing amplified warming (Chapin et al., 2005), which will cause a longer and deeper thawing of permafrost-affected soils.

As a consequence of longer and deeper thawing periods there will be an enhanced production of carbon dioxide (CO₂) through the increasing degradation of soil organic matter (SOM) (Knoblauch et al., 2013, Schuur et al., 2009, Zimov et al., 2006). On the other hand, due to higher temperatures the assimilation of CO₂ by vegetation increases because of higher plant productivity (Jia et al., 2009, Schuur et al., 2007). Belshe et al. (2013) showed that the net CO₂ uptake increased since the 1990’s in northern high-latitude tundra ecosystems, but also observed an increase in CO₂ winter emissions and therefore suggested that on an annual scale these ecosystems represent a source of atmospheric CO₂ in recent years. However, data of CO₂ fluxes from the vast Siberian permafrost landscapes are very scarce. This is critical as arctic ecosystems are highly heterogeneous and are therefore expected to show huge differences in flux strength on the microscale to macroscale (French, 1996).

Methane (CH₄) has a 34-fold higher global warming potential (GWP) as CO₂ on a 100-year timescale (Myhre et al., 2013) and is therefore an important greenhouse gas in the climate system. The arctic tundra displays a substantial source of atmospheric CH₄, but this estimate is accompanied by a large uncertainty (McGuire et al., 2012). This uncertainty can be attributed to the variability of CH₄ fluxes on spatial and temporal scales. For reliable estimates of the CH₄ source strength from arctic tundra landscapes, it is essential to perform continuous measurements across all land cover types as CH₄ emissions are
unevenly distributed in these landscapes and most likely occur during short periods of time (Marushchak et al., 2016) and may show distinct emission peaks (Mastepanov et al., 2008).

This study focussed on an arctic tundra region which holds tremendous amounts of soil organic carbon (SOC) (Zubrzycki et al., 2013). The region is a substantial sink of atmospheric CO$_2$ (Kutzbach et al., 2007b, Runkle et al., 2013) and a source of CH$_4$ (e.g. Knoblauch et al., 2015, Sachs et al., 2010, Wille et al., 2008). However, it still remains unclear which effects future climate warming will have on the C budget of permafrost-affected soils as a better understanding of the response of underlying processes of the C fluxes to warming is needed. For instance, Walz et al. (2017) have shown that the CO$_2$ release by heterotrophic respiration ($R_h$) due to decomposition of SOM will increase significantly if formerly frozen parts of the soils will thaw. Therefore, it is essential to gain a quantitative understanding of the underlying processes of CO$_2$ net ecosystem exchange (NEE) on different time scales and land cover classes. Partitioning NEE into photosynthesis, autotrophic and heterotrophic respiration is necessary to identify the response of these ecophysiological processes to a changing climate. To improve estimates of the C budgets from different tundra sites, it is necessary to investigate the CH$_4$ source strength intensely. Furthermore, a better understanding of the environmental controls on these processes is needed to improve model simulations of future C fluxes.

This study presents results of in situ chamber measurements of CO$_2$ and CH$_4$ fluxes conducted during two expeditions in a polygonal tundra in the Lena River Delta during summer 2014 and 2015.

The main objectives of this study were

1) to partition the net ecosystem exchange of CO$_2$ into the underlying processes by
   - determination of CO$_2$ fluxes and partitioning into gross primary productivity (GPP) and ecosystem respiration ($R_{eco}$) with chamber measurements
   - in situ measurements of $R_h$ fluxes over a summer period from an arctic Russian tundra site using a root-trenching approach
2) to quantify the spatio-temporal variability of the individual processes of the CO$_2$ and CH$_4$ fluxes by
1. Introduction and objectives

- calibrating flux models of GPP, $R_{eco}$ and $R_H$ as well as CH$_4$ fluxes with the measured chamber fluxes over a period of three months during the arctic growing season
- identifying dominant processes and the environmental controls governing these fluxes

The following main hypotheses were tested:

H1) The different microsites in the polygonal tundra act as sinks for atmospheric C over the measurement period

H2) The CO$_2$ fluxes differ between the microsites mainly due to contrasting hydrological conditions

H3) The contribution of $R_H$ to $R_{eco}$ during the growing season is less than 50%

H4) The contribution of $R_H$ to $R_{eco}$ show a distinct seasonality and increases by the end of the growing season

H5) The soil temperature is the environmental parameter with the highest explanatory power of the CH$_4$ fluxes

H6) Due to dry soil conditions, the polygonal rim represents a sink for CH$_4$

H7) On the ecosystem scale, the polygonal rims display a stronger net sink for atmospheric CO$_2$ than the polygonal centers
2. State of the Art

2.1. Permafrost

Permafrost is defined as soil, sediment or rock which remain below 0 °C for at least two consecutive years (van Everdingen, 2005). Almost one quarter of the total land surface of the northern hemisphere is underlain by permafrost, which are about 23 million km² (Zhang et al., 2008) (Figure 1). Based on the spatial distribution, the permafrost can be separated into continuous, discontinuous, sporadic and isolated permafrost. The permafrost depths in the continuous zone are typically ranging between 350 and 650 meters, but can reach depths of more than 1000 meters, while permafrost depths in the discontinuous zone have a much lower thickness (Schuur et al., 2008). The uppermost layer of permafrost-affected soils, the so-called active layer, thaws during the short summer period when temperatures at the surface are above 0 °C. The active layer is the layer of biological activity and hence crucial for C-cycling. The active layer depth (ALD) is important for hydrological processes, vegetation composition, quantity of SOM and all microbial degradation processes. Active layer thickness ranges from several meters in the discontinuous permafrost zone, to meters or just a few tens of centimeters in the continuous permafrost zone (Schuur et al., 2008).
2. State of the Art

2.2. Permafrost-carbon feedback

Temperatures in northern high-latitude regions have risen twice as fast as the global average with 0.6 °C per decade over the last 30 years (IPCC, 2013). This increase can lead to a positive feedback between the release of C, stored in permafrost-affected soils and the temperature changes, and this feedback is expected to accelerate (Koven et al., 2011, Schneider von Deimling et al., 2012) (Figure 2).
2. State of the Art

The arctic permafrost regions are of major interest within the context of global climate change because permafrost-affected soils store huge amounts of C (Hugelius et al., 2014, McGuire et al., 2009, Tarnocai et al., 2009). The SOC content of permafrost-affected soils is often distinctly higher than the SOC content of soils of temperate climatic zones as the decomposition of soil organic matter is inhibited due to a short growing season, low temperatures and often saturated soil water regimes as well as permanent freezing of SOC in permafrost (Hugelius et al., 2014, Ping et al., 1998, Zubrzycki et al., 2013). As a result, permafrost-affected soils acted over thousands of years as substantial sinks for atmospheric C. With rising temperatures and thawing permafrost, the formerly frozen carbon will be mobilized, resulting in an enhanced degradation of SOM (Grosse et al., 2011, Kuhry et al., 2010), which will cause an enhanced production of CO₂ (Knoblauch et al., 2013, Schuur et al., 2009). Further consequences of thawing permafrost are, among others, increased coastal erosion and subsided thermokarst pits (Jones et al., 2013, Jorgenson & Grosse, 2016).

About one quarter of the permafrost land surface is covered by tundra ecosystems (CAVM-Team, 2003). These highly heterogeneous, treeless ecosystems, whose vegetation consists primarily of grasses, sedges, herbs, shrubs, lichens and mosses, which are adapted to the extreme environmental conditions, but react sensitive to a changing climate (Kutzbach, 2006). For instance, higher temperatures will lengthen the growing season (Johnson et al., 2000), growth forms of different plant species become taller and larger (Hudson et al., 2011) and the expansion of trees and shrubs changes the plant-species composition of permafrost landscapes (Pearson et al., 2013, Tape et al., 2006). This will lead to positive feedback mechanisms on regional warming and permafrost thaw due to changes of the albedo dynamics of arctic tundra landscapes (Loranty et al., 2011) and increasing evapotranspiration (Swann et al., 2010), but can also cause lower ALD due to shrub expansion and therefore negative feedback mechanisms and higher CO₂ uptake (Blok et al., 2010). Furthermore, higher temperatures and permafrost thaw will also have impacts on microbial mineralization rates and nutrient availability in permafrost-affected soils (Beermann et al., 2015, Salmon et al., 2016, Schaeffer et al., 2013), which could increase the primary productivity and cause another negative feedback on regional warming (Natali et al., 2012, Walker et al., 2006).
2. State of the Art

Figure 2: Simplified interactions between the climate system and the carbon cycle in the Arctic. The upper layer of permafrost will thaw due to global warming which will increase the active layer thickness. This will result in the release of previously frozen carbon as CO\textsubscript{2} or CH\textsubscript{4}. This additional release of these greenhouse gases could create a positive feedback to global warming, as it will enhance the radiative forcing of the climate system (figure from Beer (2008)).

Although warming of arctic soils will enhance both the CO\textsubscript{2} uptake via GPP as well as CO\textsubscript{2} release by R\textsubscript{eco} (Mauritz et al., 2017) a further warming is suggested to decrease net CO\textsubscript{2} uptake (Parmentier et al., 2011), which is in good agreement with other studies (Schuur et al., 2013, Schuur et al., 2009). On the other hand, most recently it was suggested that on a short-term scale (decades) arctic wetlands will give a positive feedback to climate warming due to increasing CO\textsubscript{2} emissions, while this feedback could turn negative on a long-term scale (centuries) due to enhanced net C deposition (Wilson et al., 2017). However, due to additional thawed permafrost, the decomposition of formerly frozen soil organic matter could contribute substantially to future CO\textsubscript{2} release fluxes from permafrost-affected soils (Walz et al., 2017).

To estimate the effects of changing climate conditions on permafrost regions it is essential to investigate the recent carbon dynamics on the soil-atmosphere interface. Therefore, measurements of CO\textsubscript{2} and CH\textsubscript{4} fluxes are needed to improve estimates of the future sink and source function of these greenhouse gases from arctic tundra regions.
2.3. CO₂ - fluxes and processes in arctic permafrost landscapes

2.3.1. Uptake and release processes

To obtain a better understanding of the effect of further rising temperatures due to changing climate conditions on CO₂ fluxes from arctic permafrost regions, it is essential to gain a quantitative understanding of the processes underlying the CO₂ NEE, namely photosynthesis, autotrophic and heterotrophic respiration on spatial and temporal scales (Table 1). The NEE of CO₂ between the land surface and the atmosphere is composed of (1) the CO₂ uptake by plants due to photosynthesis, where atmospheric CO₂ (inorganic carbon) is converted into organic carbon (carbohydrates) and incorporated into plant biomass, described by GPP and (2) the release of CO₂ from soils and plants, described by \( R_{eco} \) (Chapin et al., 2006). Contributors to \( R_{eco} \) are autotrophic respiration by plants and roots (\( R_a \)), which includes the usage of carbohydrates for plant growth and maintenance, and \( R_H \), where litter and SOM are decomposed by soil microorganisms and used as energy supply (Luo & Zhou, 2006). In terrestrial ecosystems, sinks and sources of inorganic C are generally neglected because of their minor contribution to NEE (Elsgaard et al., 2012, Kuzyakov, 2006). Therefore, to partition NEE into the underlying fluxes, the quantification of GPP, \( R_{eco} \), \( R_a \) and \( R_H \) fluxes are required (Figure 3).
2. State of the Art

Figure 3: CO₂ and CH₄ processes in permafrost-affected soils. The most important CO₂ uptake process is photosynthesis. CO₂ release processes are plant and root respiration, together autotrophic respiration (Rₐ), and methane oxidation as well as the decomposition of soil organic matter, called heterotrophic respiration (Rₐ). CH₄ is produced in the anaerobic active layer and released by molecular diffusion, by ebullition in form of gas bubbles and through the aerenchyma of vascular plants, namely plant-mediated transport. In the aerobic active layer, CH₄ is oxidized to CO₂ by methanotrophs.

As sunlight represents the energy source of the CO₂ uptake by vegetation from the atmosphere, one of the dominant controls of GPP is the photosynthetically active radiation (PAR; Schlesinger, 2013). Beside PAR, the amount of C fixed by plants due to photosynthesis depends on temperature, plant species, water availability, air humidity, the green plant biomass and the atmospheric CO₂ concentration (Schlesinger, 2013). The temperature is also an important factor controlling ecosystem respiration fluxes because it affects almost all aspects of respiration processes (Luo & Zhou, 2006). The quality and quantity of the substrate exerts another strong influence on Rₑₑₒ (Högberg et al., 2001, Schlesinger, 2013) and soil moisture is suggested to have also a large impact on Rₑₑₒ fluxes (Xu et al., 2004). Additional factors controlling these fluxes are oxygen concentration,
nitrogen availability (carbon to nitrogen ratio (C/N)), soil texture and pH values as well as the composition of the heterotrophic community (Luo & Zhou, 2006, Schlesinger, 2013).

**Table 1**: Summary of ecosystem CO$_2$ fluxes. The CO$_2$ net ecosystem exchange (NEE) can be divided into gross primary productivity (GPP) and total ecosystem respiration (R$_{eco}$). R$_{eco}$ can be divided into autotrophic (R$_A$) and heterotrophic respiration (R$_H$). The net primary productivity (NPP) is the sum of GPP and R$_A$.

<table>
<thead>
<tr>
<th>Concept</th>
<th>Abbreviation</th>
<th>Definition</th>
</tr>
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<tbody>
<tr>
<td>Gross Primary Productivity</td>
<td>GPP</td>
<td>CO$_2$ uptake by plants during photosynthesis</td>
</tr>
<tr>
<td>Autotrophic Respiration</td>
<td>R$_A$</td>
<td>Respiratory loss of CO$_2$ by plants and roots</td>
</tr>
<tr>
<td>Heterotrophic Respiration</td>
<td>R$_H$</td>
<td>Respiratory loss of CO$_2$ by soil microbial microorganisms</td>
</tr>
<tr>
<td>Ecosystem Respiration</td>
<td>R$_{eco}$</td>
<td>R$_A$ + R$_H$</td>
</tr>
<tr>
<td>Net Ecosystem Exchange</td>
<td>NEE</td>
<td>GPP – R$_{eco}$</td>
</tr>
<tr>
<td>Net Primary Productivity</td>
<td>NPP</td>
<td>GPP – R$_A$</td>
</tr>
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</table>

**2.3.2. Partitioning ecosystem respiration**

The release of CO$_2$ from soils by R$_{eco}$ is the largest efflux of C from terrestrial ecosystems to the atmosphere (Biasi *et al.*, 2014). Therefore, changes in respiration rates can have large effects on atmospheric CO$_2$ concentration. R$_A$ can be separated into aboveground plant respiration and belowground root respiration as well as in respiration of root-derived, recent components. Due to the decomposition of SOM by heterotrophic soil organisms, R$_H$ is also called soil microbial respiration (SMR). It is challenging to separate belowground respiration fluxes into autotrophic and heterotrophic components because roots and microorganisms are closely linked within the soil (Hanson *et al.*, 2000). There is a wide spectrum of methods to partition ecosystem respiration (Kuzyakov, 2006, Subke *et al.*, 2006) and each of these approaches has its advantages and disadvantages. For different ecosystems not every approach is feasible. In ecosystems with a high root density and sparsely decomposed dead roots, like most arctic tundra ecosystems, partitioning methods based on the separation of roots are difficult to realize. Isotope labelling techniques based on $^{13}$C are principally achievable in arctic tundra landscapes (Dorrepaal *et al.*, 2009). However, because isotopic label is spread after first usage into the whole soil, it is only useful for single estimates as considerable subsequent corrections are needed for investigations on the temporal scale (Hanson *et al.*, 2000). $^{14}$C partitioning methods have been successfully applied in arctic ecosystems (Biasi *et al.*, 2014, Hicks Pries
et al., 2013, Schuur et al., 2009), but its usage on a seasonal scale to partition $R_{eco}$ with a high temporal resolution is very expensive and technically challenging.

A commonly applied method to partition belowground respiration fluxes *in situ* is root-trenching, where roots are physically isolated by digging a trench and the active plant biomass is removed (Bond-Lamberty et al., 2011, Subke et al., 2006). Despite some disturbance on the plant-soil interface, root-trenching can give accurate estimates of the rates of $R_A$ and $R_H$ (Diaz-Pines et al., 2010). Nevertheless, there are some drawbacks related to the root-trenching technique as it is associated with disturbances of the soil-root continuum (Figure 4). One of the main drawbacks is the additional decomposition of residual roots, which leads to an overestimation of $R_H$ if measurements are conducted shortly after trenching (Bond-Lamberty et al., 2011, Diaz-Pines et al., 2010, Subke et al., 2006). It is therefore preferable, that the removal of living plant biomass is conducted at least several months before the measurements are started as after this time it can be assumed that decomposition of residual roots contributes little to $R_H$ (Shurpali et al., 2008). However, the impact of other drawbacks can increase over time. The missing input of photosynthates and litter into the soil as well as turnover from roots can lead to a significant underestimation of $R_H$ (Epron et al., 2006). Furthermore, root-trenching can also have an impact on abiotic factors. For instance, Edwards and Ross-Todd (1983) have shown that root-trenching is associated with changes in soil temperatures as well as its diurnal variability. Also changes in soil moisture are ascertained to occur on trenched plots (Epron et al., 2006) which could lead to an overestimation of $R_H$. Subke et al. (2006) mentioned that the underestimation because of the lack of litter turnover and the overestimation because of soil moisture differences could cancel each other out in particular cases. On the other hand, it is assumed that hydrological conditions are re-established within days as well as soil temperature conditions are returning within weeks after root-trenching to those before the treatment (Bond-Lamberty et al., 2011, Kuzyakov, 2006). Despite the known drawbacks, it was shown that root-trenching produced similar results compared to a $^{14}$C partitioning approach in an arctic tundra ecosystem (Biasi et al., 2014).
Figure 4: Summary of disturbances accompanied by root-trenching. The disturbances can lead either to an overestimation (left) or to an underestimation (right) of the natural $R_H$ fluxes. The impact of these disturbances is both variable across different land cover classes and on temporal scale.

To date there are just a few estimates on the contribution of $R_H$ to $R_{eco}$ from tundra ecosystems during the growing season (Biasi et al., 2014, Nobrega & Grogan, 2008), and data are lacking so far for polygonal tundra ecosystems. This is critical as warming of the Arctic will influence $R_{eco}$ fluxes direct and indirect: Warming will increase decomposition of soil organic matter ($R_H$) and ensures a deeper thawing of permafrost, which will expose previously frozen SOM to decomposition (Dorrepaal et al., 2009, Schuur et al., 2011). This could cause a substantial reduction of the carbon sink function of arctic tundra ecosystems as gross ecosystem productivity is less temperature sensitive than ecosystem respiration in these ecosystems (Dorrepaal et al., 2009, Grogan & Chapin, 2000). Furthermore, warming could decrease microbial biomass C and soil moisture (Frey et al., 2008, Suseela et al., 2012) as well as increase $R_A$ due to increasing aboveground biomass, which can lead to a lower contribution of $R_H$ to $R_{eco}$ (Chen et al., 2016, Hicks Pries et al., 2015). The increase of $R_A$ and $R_H$ fluxes due to warming might be compensated by higher net primary production (Hicks Pries et al., 2013), but whether this is valid for the complete growing season and across highly heterogeneous arctic ecosystems remains uncertain.
2.3.3. Fluxes in arctic permafrost regions

To date, a set of studies focused on C balances of the arctic tundra to consider if these ecosystems act as sink or source for atmospheric CO₂ (e.g. Kutzbach et al., 2007b, Marushchak et al., 2013, Oechel et al., 2000). McGuire et al. (2012) suggested that arctic tundra regions act recently as a sink for atmospheric CO₂ on an annual basis with a total CO₂ uptake of 110 Teragram (Tg) C yr⁻¹. On the other hand, Belshe et al. (2013) found that these ecosystems act as source for atmospheric CO₂ (462 Tg C yr⁻¹). But due to limited data from Russian arctic tundra ecosystems this estimate is spatially biased. Even if this ambivalence can partly be explained by differences in the included vegetation types and a lack of data from Russian tundra within the calculation of Belshe et al. (2013), it clearly shows the need of intensified CO₂ and CH₄ flux observations in northern high-latitude ecosystems. Therefore, a better understanding of recent C dynamics in permafrost-affected soils is needed to improve estimates of future C balances of arctic tundra regions. Several studies determined NEE and R_{eco} fluxes on the microsite scale in permafrost landscapes in Alaska, Canada and Greenland (e.g. Natali et al., 2011, Nobrega & Grogan, 2008, Oechel et al., 2000, Olivas et al., 2011, Ström et al., 2012, Trucco et al., 2012), but only a few studies are available from Russian tundra ecosystems (Heikkinen et al., 2004, Kwon et al., 2016, Zamolodchikov et al., 2000). These ecosystems clearly need more attention considering their vast extent of 3 million km² (CAVM-Team, 2003), which is more than half of northern high-latitude tundra ecosystems, and their pronounced heterogeneity. Otherwise, estimates of the recent carbon balance of the arctic tundra and its future reaction on changing climate conditions are biased.

2.4. CH₄ - fluxes and processes in arctic permafrost landscapes

2.4.1. CH₄ production, oxidation and transport mechanisms

In water-saturated arctic tundra soils, CH₄ is produced as the end product of anaerobic mineralization of SOM by archaeal methanogens. Hereby, different microorganisms are converting SOM stepwise to acetate or hydrogen (H₂) and CO₂, the main reactants responsible for CH₄ production (Lai, 2009, Olefeldt et al., 2013). Within this “food-web” hydrolytic and fermenting bacteria are decomposing complex organic polymers to
alcohols, fatty acids as well as CO₂ and H₂ (Whalen, 2005). CH₄ is then produced by methanogens via two main reactions (Whalen, 2005): Hydrogenotrophic methanogens are reducing CO₂ with H₂ as electron donor:

\[ 4H_2 + CO_2 \rightarrow CH_4 + 2H_2O \] (1)

while acetotrophic methanogens produce CH₄ and CO₂ by usage of acetate as substrate:

\[ CH_3COOH \rightarrow CH_4 + CO_2 \] (2)

In northern organic soils (SOC content > 20%), Lee et al. (2012) found the hydrogenotrophic methanogenesis to be dominant, while the acetotrophic methanogenesis dominated in mineral soils. However, Vaughn et al. (2016) reported an opposing finding in polygonal tundra soils.

After production, the CH₄ is released from the soil to the atmosphere via three main pathways, namely (1) molecular diffusion, (2) plant-mediated transport and (3) ebullition (Chanton, 2005) (Figure 3).

1.) Due to the production of CH₄ in the anaerobic soil layers, there is typically a CH₄ concentration gradient and molecules are moving from higher to lower concentration following Fick’s first law (Lai, 2009). Compared to the other transport mechanisms, the molecular diffusion is rather slow, but it is important as it facilitates the contact of methanotrophic bacteria with CH₄, controlling the rate of microbial CH₄ consumption (Lai, 2009, Whalen, 2005).

2.) A major pathway for CH₄ transport is provided through the aerenchyma of some vascular plants, whereby the aerobic soil layer is bypassed (Kutzbach et al., 2004). This internal gas-space ventilation system provides aeration under anaerobic soil conditions and transports oxygen into the rhizosphere, while in the opposite direction CH₄ is transported to the atmosphere (Joabsson et al., 1999). At completely submerged arctic tundra sites, the plant-mediated transport can account for 70 to 90% of the total CH₄ emissions from the soil to the atmosphere (Knoblauch et al., 2015).

3.) Another pathway in water-saturated soils is the sudden release of CH₄ from the soil by ebullition, which can contribute substantially to total CH₄ fluxes (Tokida et
al., 2007). These gas bubbles are formed when the partial pressure of all dissolved gases in solution exceeds the hydrostatic pressure in the soil (Lai, 2009).

In the aerobic soil layers, parts of the produced CH$_4$ is consumed as energy source for growth and maintenance by methanotrophs (Hanson & Hanson, 1996). Hereby, CH$_4$ is sequentially oxidized to methanol, formaldehyde, formate and eventually CO$_2$ (Whalen, 2005):

$$\text{CH}_4 \rightarrow \text{CH}_3\text{OH} \rightarrow \text{HCHO} \rightarrow \text{HCOOH} \rightarrow \text{CO}_2$$

In peatlands, the highest methanotrophic activity is located at the aerobic-anaerobic interface close to the water table as these interface provides the optimal ratio of the substrates, oxygen and CH$_4$ (Dedysh, 2002, Knoblauch et al., 2015, Lai, 2009, Preuss et al., 2013).

2.4.2. Environmental controls on CH$_4$ fluxes

CH$_4$ emissions from arctic wetlands are a product of the balance between methanogenesis and oxidation, which is substantially influenced by the presence of oxygen, one of the most important factors controlling CH$_4$ emissions (Whalen, 2005). If anaerobic conditions exist, the substrate supply is the limiting factor of methanogenesis (Ström et al., 2012, Yavitt & Lang, 1990). Similar to CO$_2$ respiration fluxes, the (soil) temperature is one of the main factors controlling both methanogenesis as well as methane oxidation (Knoblauch et al., 2008, Lai, 2009). Vegetation composition can influence CH$_4$ oxidation rates as there is evidence for symbiotic-like relations of methanotrophs with moss species (Liebner et al., 2011). Wagner et al. (1999) found soil texture to be an important parameter influencing CH$_4$ production. Also, the soil and porewater pH has an impact on CH$_4$ production and oxidation mechanisms, but the limiting and the optimal pH values are varying with different ecosystems (Whalen, 2005). Furthermore, CH$_4$ emissions are heavily affected by fluctuations of the water table depth, which can be explained with changes of the thicknesses of potential CH$_4$ production and oxidation zones (Lai, 2009).

The water table depth is one of the main predictors of CH$_4$ emissions as it affects significantly the production and oxidation rates (Marushchak et al., 2016, McEwing et al., 2015) and is mostly negatively correlated with the CH$_4$ fluxes (Lai, 2009). Another
important predictor is the vegetation composition (Davidson et al., 2016); especially the presence of sedges can have large impacts on the rate of CH$_4$ emissions as they promote plant-mediated transport (King et al., 1998, Knoblauch et al., 2015, Kutzbach et al., 2004, Olefeldt et al., 2013). CH$_4$ emissions data from Eddy covariance (EC) measurements showed near-surface turbulence to be one of the main drivers of CH$_4$ emissions (Sachs et al., 2008, Wille et al., 2008). If the ALD and permafrost degradation can serve as parameter influencing CH$_4$ emissions is highly uncertain (Olefeldt et al., 2013, Prater et al., 2007, van Huissteden et al., 2005). Various studies have identified the temperature as one of the main controls of CH$_4$ emissions (e.g. Mastepanov et al., 2013, Sachs et al., 2010, Schneider et al., 2016, Wille et al., 2008). This relationship is based on the temperature dependence of microbial activity (Conrad, 1996, Wagner et al., 2007).

**2.4.3. CH$_4$ flux estimates from arctic permafrost landscapes**

Due to often water-saturated and anaerobic conditions, the arctic tundra displays a significant source of atmospheric CH$_4$ with about 19 Teragram (Tg) C yr$^{-1}$, ranging from 9 to 35 Tg C yr$^{-1}$ (McGuire et al., 2012). Considering worldwide CH$_4$ emissions from wetlands between 133 and 213 Tg C yr$^{-1}$ (Ciais et al., 2013) and total global emissions of 550 Tg C yr$^{-1}$ (Dlugokencky et al., 2011), arctic tundra regions are a substantial source of atmospheric CH$_4$. Therefore, changes in the CH$_4$ source strength of arctic tundra regions in the future can have large impacts on the global greenhouse gas budgets. For instance, rainfall is predicted to increase in high-latitudes (Christensen et al., 2013), and as a result of that CH$_4$ emissions could increase as anaerobic soil conditions are expected to expand with higher precipitation rates (Sachs et al., 2010). Furthermore, thaw-induced subsidence is predicted to increase CH$_4$ emissions (Johnston et al., 2014, Olefeldt et al., 2013). On the other hand, permafrost thaw causes morphological changes of arctic landscapes, resulting in increased drainage (Godin et al., 2014) and arctic warming causes more evaporation which leads to less summer inundation (Koven et al., 2011). Both processes can cause a reduction of the CH$_4$ source strength from arctic tundra regions. Furthermore, non-saturated soils in high-latitudes can also act as substantial sinks for atmospheric methane (Jørgensen et al., 2015, Zhu et al., 2014) and also water-saturated sites can act as sinks when exposed to high radiation (Liebner et al., 2011). The CH$_4$ sink
strength of non-saturated soils is suggested to increase in future simultaneously to warmer soil conditions (Jørgensen et al., 2015).

The annual \( \text{CH}_4 \) fluxes in high-latitudes are suggested to be dominated by fluxes during the short growing season (e.g. Whalen & Reeburgh, 1992, Wille et al., 2008), but recently, the emissions from the cold season (September to May) are found to contribute significantly to annual \( \text{CH}_4 \) budgets (Zona et al., 2016). Nevertheless, a detailed understanding on the environmental controls and sensitivities of growing season \( \text{CH}_4 \) emissions is needed to assess the impact of future changing climate conditions on \( \text{CH}_4 \) fluxes.

2.5. Chamber measurement technique

Closed chamber methods are widely used to quantify emissions of trace gases (Livingston et al., 2005) as they allow a good spatial coverage. They are low in cost, simple to use and can be applied in remote, low-stature canopies like arctic tundra regions (Kutzbach et al., 2007a). However, the use of closed chambers is related to a set of potential errors, including (1) the disturbance of soils during collar installation, (2) the disturbance of pressure gradients during chamber deployment, (3) changes of turbulence strength, (4) inaccuracies in the calculation of headspace volume and leakage closure as well as (5) the alteration of the natural gas concentration gradient inside the chamber headspace (Davidson et al., 2002, Hutchinson & Livingston, 2001, Kutzbach et al., 2007a, Lai et al., 2012, Redeker et al., 2015). All of these potential errors can lead to a serious bias of the measured fluxes and the latter leads to non-linear concentration-over-time curves of chamber headspace air (Kutzbach et al., 2007a, Pedersen et al., 2010). A non-linear flux evaluation can also have methodological reasons. For instance, if the chamber is not equipped with large openings on top, air is pressed into the soil during chamber deployment on the soil, which can lead to initial pressure shocks (Christiansen et al., 2011, Schneider et al., 2011). These initial pressure shocks can cause a sudden release of gases from the soil, which is perturbing the concentration-over-time curve of a chamber measurement and lead to substantial overestimation of the fluxes. However, Görres et al. (2014) found out that the flux calculation with a linear regression model is more appropriate to describe \( \text{CO}_2 \) fluxes even if the concentration-over-time curves are
non-linear. Furthermore, chambers have to be equipped with a fan to allow continuous headspace mixing, and should be in an adequate size as both can lead flux underestimation (Christiansen et al., 2011, Hutchinson & Livingston, 2001, Pihlatie et al., 2013). However, the use of fans to mix chamber headspace air is part of current discussion (Redeker et al., 2015). Some studies showed that realistic flux estimates of trace gas emissions can be achieved only in chambers where the air was mixed by a fan (Christiansen et al., 2011, Denmead, 2008), but some authors suggested that chambers that are equipped with fans give unreliable readings (Davidson et al., 2002). Another important factor to keep potential errors to a minimum is the exclusion of the data from the very beginning of the chamber measurement to allow steady headspace mixing and prevent that disturbances caused by chamber deployment are included into the flux calculation (Christiansen et al., 2011, Koskinen et al., 2014). This exclusion is also important for measuring R\textsubscript{eco} as CO\textsubscript{2} uptake can continue for a short time period after PAR is blocked in dark chambers (Laisk et al., 1984).

If all these potential errors are kept to a minimum, the use of closed chambers is appropriate for the observation of the spatial and temporal variability of CO\textsubscript{2} and CH\textsubscript{4} fluxes. Arctic polygonal tundra landscapes are on the microscale (1 m to 10 m) highly heterogeneous landscapes (French, 1996). Therefore, measurements on this microscale are necessary as solely they can give insights into the processes controlling the CO\textsubscript{2} and CH\textsubscript{4} fluxes and the future reactions of these fluxes on changing climate conditions. Chamber measurements can provide reliable estimates of gas fluxes from this microscale. These estimates can further be used to reproduce the gas fluxes over complete seasons by developing flux models based on the measured fluxes. The models provide crucial information on the dependence of the gas fluxes from environmental parameters and they can give reliable estimates of cumulative fluxes. Therefore, they can be used for the quantitative contribution of single fluxes and microsites to the arctic C cycle.
3. Study area

3.1. Lena River Delta

The Siberian Lena River Delta (72.0 – 73.8°N, 122.0 – 129.5°E) is the third-largest river delta worldwide and the largest delta in the Arctic and covers a surface area of 29,000 km² (Schneider et al., 2009) (Figure 6). This fan-shaped delta is characterized by a network of rivers and channels with more than 1,500 islands of different sizes. The Lena River Delta can be geomorphologically divided into three main river terraces of different ages and the flood-plain levels (Grigoriev, 1993 in Schwamborn et al., 2002). The first terrace includes active flood-plains, large thermokarst lakes and ice-wedge polygonal tundra. This youngest terrace, formed in the Middle Holocene, occupies about two-third of the total delta area, primarily occurring in the central and eastern parts of the delta (Bolshiyanov et al., 2015, Schwamborn et al., 2002). The second terrace occurs in the northwestern part of the delta and is characterized by sandy sediments with a low ice content; it was formed between the Late Pleistocene and the Early Holocene and covers about 23% of the delta (Schneider et al., 2009). The oldest terrace consists of fine-grained, organic- and ice-rich sediments and is an erosional remnant of a Late Pleistocene plain, characterized by polygonal ground and thermokarst processes (Boike et al., 2013).

The delta is located in the continuous permafrost zone with permafrost depths of about 500 to 600 m (Grigoriev, 1960) and relatively low temperatures of -7.8 °C at 1.7 m soil depth (Boike et al., 2013). A mean annual permafrost temperature of -8.6 °C at 10.7 m depth demonstrates that the Lena River Delta is one of the coldest permafrost regions on the earth (Romanovsky et al., 2010).

The climate in the Lena River Delta is arctic-continental and characterized by both low temperatures and precipitation (Boike et al., 2008). Despite comparatively low precipitation, the climate of the delta has to be considered as humid because of low evaporation rates due to low temperatures (Kutzbach, 2006). At the weather station in Tiksi (about 120 km southeast of the study site in the delta), the mean annual air temperature over a 30-year period (1981-2010) was -12.8 °C and the mean annual precipitation 321 mm (Figure 5). The mean temperatures of the warmest (August: 7.8 °C)
3. Study area

and the coldest month (January: -30.2 °C) highlight the extreme climatic contrast between the seasons, typical for continental polar regions. The summer growing season, defined in this study as the period with consecutive positive daily average air temperatures, usually lasts about three months, from mid-June until mid-September. The mean summer precipitation at the study site during the period 1998-2011 was 125 mm, ranging from 52 mm to 199 mm (Boike et al., 2013), which is about 40% of the annual precipitation. The central delta region experiences rapidly changing weather conditions during summer by advection of cold and moist or warm and dry air from the Arctic Ocean or continental Siberia, respectively (Runkle et al., 2013). Polar day lasts from 7 May until 8 August and polar night lasts from 15 November to 28 January. River ice break-up and snowmelt starts usually in the first half of June.

Figure 5: Climate chart of Tiksi over the period 1980-2011. The reference site is located ~120 km southeast of the study site. The precipitation and temperature data are shown as monthly mean values.

3.2. Samoylov Island

The study area is located on Samoylov Island in one of the main river channels, the Olenyokskaya Channel, in the southern central Lena River Delta (72°22′N, 126°28′E; Figure 6). During the past two decades, a number of studies with focus on geomorphology, microbiology, soil science as well as gas and energy exchange were
conducted on Samoylov Island (e.g. Boike et al., 2008, Kutzbach et al., 2007b, Muster et al., 2012, Sachs et al., 2010, Schwamborn et al., 2002). Samoylov Island covers an area of around 5 km². The island can be geomorphologically divided in two parts: In the western part of the island is an active flood-plain (2 km²) with elevations from 1 to 5 m above sea level (a.s.l) which is occasionally flooded during river break-up. The study site is located in the eastern part of the island in the wet ice-wedge polygonal tundra on a Late Holocene river terrace (3 km²) with elevations from 10 to 16 m a.s.l. This River terrace is only partly flooded during extreme flooding events.

Figure 6: The study site on Samoylov Island, Lena River Delta in Northeastern Siberia (72°22’N, 126°28’E). (Images: Lena River delta - left: NASA Landsat program: Lena River Delta in Landsat 7; available at: http://earthobservatory.nasa.gov/IOTD/view.php?id=2704, 2002; Samoylov Island - middle: Boike et al. (2012); Study site - right: Boike et al. (2015))

Wet polygonal tundra landscapes are typical for the Late Holocene River Terrace in the Lena River Delta (Figure 7). Polygonal structures are developing due to thermal contraction in the winter season, which leads to cracks of the surface followed by ice-wedge growth when melting water freezes in these cracks. This process produces a honeycomb-like surface structure with high- and low-centered polygons as well as polygonal ponds. In low-center polygons (hereinafter ‘polygonal center’) drainage is impeded due to the underlying permafrost and as a result of that, anaerobic accumulation of organic material is feasible (Wagner et al., 2003). In contrast to that, in the soils of the higher situated polygonal rims, which are surrounding the polygonal centers, less SOM is accumulated due to oxic conditions in the upper soil part. The elevation differences between wet polygonal centers and comparatively dry polygonal rims is about 0.5 m.
3. Study area

Figure 7: The polygonal tundra on Samoylov Island. Polygons are formed due to thermal contraction in winter that produces cracks where ice-wedges can grow when melting water freezes later inside these cracks. The study site (white circle), the soil measurement station (upper left) and the eddy covariance system (bottom left) are connected with boardwalks to avoid disturbance. Image from Boike et al. (2015)

On both parts of the island an eddy covariance (EC) system for measurements of energy, wind and gas fluxes was installed (Kutzbach et al., 2007b, Runkle et al., 2013, Wille et al., 2008). The study site in the current work is located within the footprint area of the EC system on the river terrace. Surface classification of aerial images had shown that polygonal rims (dry tundra) represent about 50% while depressed polygonal centers (wet tundra) represent approximately 40% and open water bodies represent around 10% of the EC footprint area (Muster et al., 2012). With more than 25 kg m\(^{-2}\) SOC in the uppermost meter these landscapes contain a significant pool of SOM (Zubrzycki et al., 2013).

In this study, two different microsites were investigated (Figure 8): a wet-depressed polygonal center and its surrounding elevated polygonal rim (72°22′442 N; 126°29′828 E). At this polygon, the maximum ALD is deeper at the polygonal center with about 40 cm
3. Study area

compared to the polygonal rim with about 30 cm. Nutrient input into this rather nutrient-limited ecosystem comes from upward migration of fluvial sediments due to cryoturbation at the polygonal rims and maybe from aeolian deposits. According to WRB (2014) the soils at the polygonal center are typically classified as Histic Cryosols with water tables close to the soil surface. The polygonal rim soils are characterized by cryoturbation, a rearrangement of soil material due to freeze-thaw processes. Therefore, the soils at the polygonal rim are classified typically as Turbic Glacic Cryosols with a water table just a few centimeters above the permafrost table. High contents of SOM (> 10%) are usually occurring in the first 15 cm within the soils of the polygonal rim and are underlain by cryoturbated mineral horizons, while high SOM contents at the polygonal center are common in the entire active layer (40 cm). The vegetation of the polygonal rim is dominated by the moss species Hylocomium splendens as well as Dryas punctata, Astragalus frigidus and lichens (Peltigera spp.), whereas vegetation of the polygonal centers is dominated by the hydrophilic sedge Carex aquatilis and mosses (Drepanocladus revolvens, Meesia triqueta, Scorpidium scorpioides).

Figure 8: Polygonal rim and center at the study site. The typical vegetation of polygonal rims (left) is moss-dominated with small amounts of lichens and various vascular plants, while polygonal centers (right) can be considered as sedge-moss-dominated.
4. Methods

4.1. Meteorological data

Meteorological variables were recorded in 30 min intervals at the nearby EC system and an adjacent meteorological station, 40 m southwest from the study site. An ultrasonic anemometer (Solent R3, Gill Instruments Ltd, UK) measured wind velocity components in three dimensions. Further data on relative humidity and air temperature (MP103A, ROTRONIC AG, Switzerland), air pressure (RPT410F, Druck Messtechnik GmbH, Germany) and photosynthetic active radiation (PAR; wavelength: 400 – 700 nanometer; QS2, Delta-T Devices Ltd., UK) as well as the incoming and reflected components of shortwave and longwave radiation, respectively (CNR 1, Kipp and Zonen, Netherlands) were collected. The radiative surface temperature \( T_{\text{surf}} \) was calculated after Kutzbach \textit{et al.} (2007b) using the formula

\[
T_{\text{surf}} = \left( \frac{L_{\uparrow}B}{\varepsilon 5.67 \times 10^{-8}} \right)^{1/4}
\]

where \( L_{\uparrow}B \) is the upward infrared radiation and the emissivity \( \varepsilon \) was assumed to be 0.98.

4.2. Soil, vegetation and porewater analysis

Soil analysis was done in July 2014. At the polygonal rim a total of six soil samples were taken using steel rings between the soil surface and the frozen ground with a depth of 12-16 cm and 6 cm in inner diameter. These samples were separated into an organic-rich and a mineral soil layer based on soil texture. Between these soil layers a transition layer was determined, developed due to cryoturbation processes. At the polygonal center one soil sample was taken from the surface to the frozen ground with a spade and separated vertically into three replicates. These samples were separated into horizontal layers after SOM decomposition status. Living plant biomass from the top was removed from all soil samples. For further analysis, the samples were dried at 70 °C and milled. Soil organic carbon contents were measured with a liquiToc II coupled with a solids module (Elementar Analysysysteme GmbH, Hanau, Germany) and total contents of carbon and nitrogen (N) were measured with an Element analyzer (VarioMAX cube, Elementar Analysysysteme GmbH, Hanau, Germany). These values were used to calculate the C/N
ratio. The weight difference between wet and dried (105 °C) soil samples were used to calculate soil water contents. Soil pH values were measured in a suspension of 5 g soil in 12.5 mL distilled water (CG820, Schott AG, Mainz, Germany).

Furthermore, to compare CO₂ and CH₄ fluxes with vegetation community structure each chamber collar (50 x 50 cm) was divided into 10 x 10 cm subplots. In four of these subplots the plant species were recorded and their abundance (% of the surface cover) was determined. This procedure was conducted at every single measurement plot in mid-July and at the end of August in 2015.

Depth profiles of pH values at the water-saturated polygonal center were analyzed with a field pH meter (340i, WTW, Xylem Inc., Weilheim, Germany). Therefore, porewater was extracted from the center using a steel rod with small openings at the bottom and an attached syringe at the top. At first, the rod and the syringe were flushed with water from the required depth. In a second step, 150-200 mL of porewater were extracted from the soil. Porewater samples were taken in 2015 from 5, 10, 15, 20, 25, 30 and 35 cm soil depth in weekly intervals. Samples from the latter were taken when the active layer was thawed to this depth (first on 26 August). Description of pH values were made following the protocol of Schoeneberger et al. (2012).

4.3. Chamber measurements

4.3.1. Light and dark chamber measurements

A total of eight PVC chamber collars (50 x 50 cm), four at each microsite, were permanently installed in July 2014 at the sampling site for CO₂ as well as CH₄ flux measurements (Figure 9). Both microsites were equipped with boardwalks to avoid disturbance. The chamber collars had boreholes with 5 cm in diameter on each of the four sides in 15 cm soil depth to allow lateral water movement and were inserted 20 cm deep into the active layer. The collars were equipped with an U-shaped frame filled with water to avoid gas exchange between the chamber headspace and ambient air. The chamber used for CO₂ and CH₄ flux measurements was made of clear plexiglas (Plexiglas SunActive GS, Evonik Industries AG, Germany) with dimensions of 50 x 50 x 50 cm and a thickness of 4 mm. For Rₑₑₒ measurements the chamber was covered with an optically opaque box.
4. Methods

(dark chamber measurements). The chamber was equipped with a fan for continuous mixing of headspace air (axial fan, 12V/DC, Conrad Electronic SE, Germany). Furthermore, a PAR sensor (SKP212, Skye Instruments Ltd., UK) and a temperature probe (107 Thermistor probe, Campbell Scientific Ltd., USA) were installed inside the chamber. Including the volume inside the chamber collars, the chamber enclosed a volume between 124-143 L. In 5 cm soil depth, soil temperature and volumetric water content (VWC – GS3, Decagon Devices, Inc., USA) were measured during every single chamber measurement right beside the chamber collar. A Diver (Schlumberger Ltd., USA) was installed at the polygonal center to measure water table (WT) fluctuations in 15 min intervals. To prevent pressure induced emission shocks during chamber closure, two openings (3 cm in diameter) were installed at the top of the chamber that were left open during the chamber placement and closed immediately after placing the chamber on the collars. On every measurement day soil temperature between the surface and the frozen ground in 5 cm intervals and ALD were measured at both microsites. CO₂, CH₄ and water vapor (H₂O) concentrations in the chamber headspace were measured with an Ultra-Portable Gas Analyzer (UGGA 30-p, Los Gatos Research, USA). The chamber headspace was pumped in a closed loop via transparent PUN tubes (inner diameter 4 mm, each 10 m length) through the analyzer with a flowrate of 200 mL min⁻¹. Gas concentrations of CO₂, CH₄ and water vapor were logged together with PAR as well as soil and air temperature with a frequency of 1 hertz (Hz) on a data logger (CR800series, Campbell Scientific Ltd., USA). Each chamber measurement was restricted to 120 seconds to minimize warming effects inside the chamber.
4. Methods

**Figure 9**: Diagram of the study site and the installed measurement plots. An elevated polygonal rim surrounds a wet-depressed center. Overall, 20 PVC collars, 10 on each microsite, were installed in 2014 (16) and 2015 (4) for CH₄ and CO₂ measurements and a boardwalk to avoid disturbances.

A total of 2,020 chamber measurements were conducted in summer 2015 from 11 July until 22 September. Except the weeks from 2 to 9 August (shift change) and 17 to 24 August (storm event) measurements were done at least every third day between 6 am and 9 pm (local time). Between measurements, atmospheric background concentration were achieved inside the chamber by holding the open-bottom of the chamber to the wind. Two consecutive measurements were performed at each collar: First, NEE and CH₄ (n = 679) were measured simultaneously with the transparent chamber followed by an \( R_{\text{eco}} \) measurement (n = 679) with the dark chamber. The four collars of one microsite were measured in a row before moving to the other microsite. The GPP fluxes were calculated from the sum of the measured \( R_{\text{eco}} \) and NEE fluxes.
Furthermore, 860 chamber measurements have been made from 23 July to 20 August in 2014 to observe CO₂ and CH₄ fluxes at the polygonal center and rim. The measurement procedure was identical to those of the chamber measurements in 2015.

4.3.2. Root-trenching approach

For Rₜ measurements the root-trenching method was applied at both microsites (Figure 10). By inserting PVC collars 20 cm deep into the soil, which is below the main rooting zone, lateral roots were cut off. All living plant biomass including mosses inside the collars was removed carefully in 2014. To prevent re-growth, the living plant biomass was removed periodically during the complete measurement period. This manipulation causes the die-off of roots (excludes Rₐ) and hence Rₜ fluxes can be determined. In total, the living plant biomass was removed from the inside of eight collars, four at each microsite. Repeatedly light and dark chamber measurements were made consecutive at the same plot to ensure the absence of photosynthesis. Rₜ fluxes (n = 662) were measured in 2015 during the same periods and with the same measurement interval as NEE, CH₄ and Rₑₑₒ measurements on unaltered collars. In 2014, 226 chamber measurements on the altered plots were made.

To test if Rₜ fluxes are related to artefacts due to the root-trenching approach, four additional (two on each microsite) PVC collars were installed in 2015 and the living plant biomass was removed. The sampling protocol was the same to those plots where the root-trenching was applied in 2014. 302 Rₜ flux measurements were made on these newly installed plots. The differences of the mean Rₜ fluxes of every single plot that were trenched in 2014 and 2015 were analyzed using a student’s t-test. Rₐ fluxes were calculated by subtracting the measured Rₜ fluxes from the measured Rₑₑₒ fluxes that were measured simultaneously. The calculated Rₐ fluxes were used to calculate the net primary productivity fluxes (NPP), which is the sum of GPP and Rₐ fluxes.
4. Methods

Figure 10: The root-trenching approach at the polygonal center. The living plant biomass was removed from the measurement plots (left) to measure $R_{vi}$ fluxes, while the vegetation at other measurement plots was left intact (right) to measure CH$_4$, NEE and $R_{eco}$ fluxes at the same microsite. The root-trenching was conducted in 2014. In 2015 the method was applied at four additional measurement plots to observe if the $R_{vi}$ fluxes are related to artefacts.

4.4. Flux calculation

CO$_2$ and CH$_4$ fluxes (in microgram (µg) CO$_2$/CH$_4$ m$^{-2}$ s$^{-1}$) were calculated in MATLAB® R2015a (The MathWorks Inc., Natick, MA, 2000) with a routine which calculates flux rates with linear and different non-linear regression models and provides information criteria to find the regression that fits best (Eckhardt & Kutzbach, 2016). The regression models are described in detail in Kutzbach et al. (2007a). For flux rate calculations, a linear regression model and a Taylor power series expansion of the exponential model were used. The results of the power series expansion model are practically identical to those obtained from the exponential function, but the power series expansion model directly estimates the initial slope of the flux curve which results in lower error estimates for the initial slope. To determine if the fluxes are best represented by the linear or the power series expansion function the Akaike Information Criterion for finite sample sizes ($AIC_c$) was used (Burnham & Anderson, 2004) and the adjusted non-linear coefficient of determination ($R^2_{adj}$). It revealed that for the CO$_2$ fluxes, the linear model provided a better fit in most cases, with a lower $AIC_c$ value in 71% and a higher $R^2_{adj}$ in 53% of cases. For the CH$_4$ fluxes the linear regression model provided better fit with a lower $AIC_c$ in 88% and a higher $R^2_{adj}$ in 67% of cases.
The CO$_2$ and CH$_4$ fluxes were calculated using the following equation:

\[ F_{\text{CO}_2, \text{CH}_4} = \frac{M_{\text{CO}_2, \text{CH}_4}}{R} \times \frac{V}{A} \times \frac{P_{\text{atm}}}{T_{ch}} \times \frac{d\text{CO}_2, \text{CH}_4}{dt} \]

where \( M_{\text{CO}_2, \text{CH}_4} \) is the molar mass of CO$_2$ and CH$_4$, respectively (g mol$^{-1}$), \( R \) is the ideal gas constant (J K$^{-1}$ mol$^{-1}$), \( V \) is the chamber headspace volume (m$^3$), \( A \) is the chamber surface area (m$^2$), \( P_{\text{atm}} \) is the atmospheric pressure (Pa) and \( T_{ch} \) is the temperature inside the chamber during the measurement in Kelvin. The last term \( \frac{d\text{CO}_2, \text{CH}_4}{dt} \) describes the concentration evolution of the investigated gas in the chamber headspace over time.

Due to possible perturbation of the concentration-over-time curve during chamber setting, the first 30 seconds of each 2-minute measurement period were discarded and the remaining 90 data points were used for flux calculation. According to the manufacturer, the precision of the Gas Analyzer with 1 s signal filtering is < 0.3 parts per million (ppm) for CO$_2$ and < 0.005 ppm for CH$_4$. Typically, the root mean square error (RMSE) of chamber measurements and model fitting did not exceed these values and therefore higher RMSE values indicate failed model fitting or disturbed chamber measurements. Therefore, if RMSE exceed 0.3 ppm for CO$_2$ fluxes and 0.005 ppm for CH$_4$ fluxes, the flux curve was re-inspected. If irregularities could be removed by adjusting the measurement period, the flux curve was re-calculated and if not the dataset was discarded. Variation of PAR during chamber employment due to shifts in cloud cover leads to perturbation of the CO$_2$ concentration-over-time curves (Schneider et al., 2011). These perturbed curves show distinct autocorrelation of the residuals of the fit function. Therefore, they were filtered out using a threshold for residual autocorrelation indicated by the Durbin-Watson test (Durbin & Watson, 1950). Overall, in 2015 about 3% (\( n = 47 \)) of the CO$_2$ flux measurements (NEE, \( R_{\text{eco}} \) and \( R_{\text{H}} \) measurements) were discarded from the dataset, because they did not meet the mentioned quality criteria. For the same reasons, almost 4% (\( n = 26 \)) of the CH$_4$ flux measurements were removed. From the 2014 dataset a total of 37 chamber measurements (~4%) were discarded as they exceeded the thresholds of RMSE and/or Durbin-Watson test.
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As the chamber in 2014 was not equipped with large openings until 3 August, the CH$_4$ flux measurements cannot be used from measurements made between 22 July and 2 August. Missing openings at the top of the chamber can lead to a substantial overestimation of CH$_4$ emissions due to caused initial pressure shocks (Christiansen et al., 2011). Therefore, 180 chamber measurements were excluded from the CH$_4$ dataset (~45%).

4.5. Modelling carbon fluxes

4.5.1. CO$_2$ model

There are two different sign conventions for NEE: the ecological sign convention defines a positive NEE as net uptake of CO$_2$ by the plant-soil ecosystem, whereas the atmospheric sign convention defines it as a net release of CO$_2$ from the soil to the atmosphere. In this study the atmospheric sign convention is used where a positive NEE defines a net release of CO$_2$ from the soil to the atmosphere.

GPP, R$_{eco}$ and R$_H$ fluxes from 2015 were modeled separately. The chamber measurements from 2014 were not used for modelling as the dataset is too small in quantity as well as on temporal scale. Examples of the modeled GPP and R$_{eco}$ fluxes are given in Figure 11. To calibrate the models, the measured GPP, R$_{eco}$ and R$_H$ fluxes were fitted to the used functions for the flux models. The resulting fitting parameters were used to reproduce the fluxes over the complete measurement period. The modelling was done with a moving window of 15 days during the measurement period. If less than eight chamber measurements were performed in these 15 days, the moving window was extended to 19 days. Each of the four measurement plots per microsite were modeled separately and the summed fluxes were used to analyze differences between both microsites with a student’s t-test. Subsequently, the modeled fluxes for each measurement plot were averaged for each microsite. The empirical $Q_{10}$ model (van’t Hoff, 1898) was fitted to the measured R$_{eco}$ and R$_H$ fluxes:

$$R_{eco,H} = R_{base} \times \frac{T_{surf,soil-T_{ref}}}{\gamma} \quad (6)$$

where the fit parameter $R_{base}$ is the respiration at the reference temperature $T_{ref}$. $T_{ref}$ (15 °C) and $\gamma$ (10 °C) were held constant according to Mahecha et al. (2010). $Q_{10}$ is a fit parameter indicating the ecosystem sensitivity to a 10 °C change in temperature. For this
study a fixed $Q_{10}$ of 1.52 was used, which represents the seasonal mean value of the bulk partitioning model for the EC footprint area (Runkle et al., 2013). Air temperature ($T_a$), surface temperature ($T_{surf}$) and soil temperature ($T_{soil}$) measured at a depth of 5 cm were tested as input variables. Since it has been shown that the water table and volumetric water content are important environmental factors in explaining $R_{eco}$ and $R_H$ fluxes (Biasi et al., 2014, Schneider et al., 2011), the respiration fluxes from the polygonal center were also related to WT and from the polygonal rim to VWC:

$$R_{eco,H} = R_{base} \times \frac{T_{a,surf,soil}-T_{ref}}{\gamma} \times \exp^{(WT,VWC \times c)} \quad (7)$$

The additional fitting parameter $c$ was set to -0.11 when the model was calibrated to fluxes from the center and to -0.5 when the model was calibrated to fluxes from the rim. These values constitute the medians of this parameter obtained from test runs when $c$ was held variable.

The modeling was done with MATLAB® R2015a (The MathWorks Inc., Natick, MA, 2000). The model parameters were estimated by nonlinear least-squares fitting (nlinfit function) and the uncertainty of the parameters were determined by calculating the 95% confidence intervals using the nlparci function. The selection of the best performing $R_{eco}$ and $R_H$ model was based on comparing the $R^2_{adj}$. The selected model was chosen as model for all measurement plots, even though for a single plot another model had a better qualifying parameter.

For an estimate of GPP, the $R_{eco}$ fluxes, which were measured directly after the NEE fluxes were subtracted from the NEE fluxes. GPP fluxes are reproduced with the rectangular hyperbola function as a function of PAR (in µmol m$^{-2}$ s$^{-1}$):

$$GPP = -\frac{P_{max} \times \alpha \times PAR}{P_{max} + \alpha \times PAR} \quad (8)$$

The fit parameters $P_{max}$ and $\alpha$ are the maximum canopy photosynthetic potential (hypothetical maximum of $P_{max}$ at infinite PAR) and the initial canopy quantum efficiency (initial slope of the $P_{max}$ - PAR curve at PAR = 0), respectively. The values for $\alpha$ were obtained from modelling the CO$_2$ fluxes with EC data (Kutzbach et al., unpublished). From the determined values when $\alpha$ was variable, a function was formulated with specific values for every day of the growing season using the following function:
\[ \alpha = b \times e^{\frac{-abs((x-c)\times d)}{2\times e^2}} + f \]

where \( b = 0.042, c = 209.5, d = 2, e = 25.51, f = 0.008 \) and \( x \) = day of year 2015. Afterwards, these “semi-fixed” values (variable on daily basis) were used to reproduce GPP fluxes from chamber measurements over the complete measurement period.

Afterwards, these “semi-fixed” values (variable on daily basis) were used to reproduce GPP fluxes from chamber measurements over the complete measurement period. Figure 11: Examples of the models for the reproduction of GPP and \( R_{eco} \) fluxes. The relationship between PAR and GPP fluxes (left) is characterized by a light response curve (dotted line) and chamber measurements are shown as crosses. The temperature sensitivity of the \( R_{eco} \) fluxes (right) is exponential (dotted line). Stars display \( R_{eco} \) chamber measurements.

To consider the different phases of plant development throughout the growing season, three other GPP models were tested: (1) a rectangular hyperbola function including the normalized difference vegetation index (NDVI) as well as the ratio vegetation index (RVI) as input variable, (2) a sigmoidal model as well as a light response model including the leaf area index (LAI) and (3) a simple logistic function (Görres et al., 2014, Marushchak et al., 2013, Rodeghiero & Cescatti, 2005, Wohlfahrt et al., 2010). The values of the NDVI and LAI were obtained from satellite images (ORNL DAAC, 2014). The values of the RVI were calculated using the following formula (Wilson & Meyers, 2007):

\[ RVI = \frac{(NIR_{out}/NIR_{in})}{(PAR_{out}/PAR_{in})} \]

where \( NIR_{out} \) and \( NIR_{in} \) (in W m\(^{-2}\)) are the outgoing and incoming near infrared radiation and \( PAR_{out} \) and \( PAR_{in} \) (in \( \mu \)mol m\(^{-2}\) s\(^{-1}\)) are the outgoing and incoming PAR, respectively. Unfortunately, all these models produced unrealistic fitting parameter and/or fluxes and where therefore not used in this study (Table 2).
Table 2: Failed GPP, R\textsubscript{eco} and R\textsubscript{H} models. All listed functions produced unrealistic fitting parameter and/or fluxes.

<table>
<thead>
<tr>
<th>ID</th>
<th>Model formula</th>
<th>Remarks</th>
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<tbody>
<tr>
<td>A</td>
<td>( GPP = \frac{\beta \times RVI \times \alpha \times PAR}{\beta \times RVI + \alpha \times PAR} )</td>
<td>Modification of rectangular hyperbola model (after Görrres \textit{et al.} (2014)); Input variables are RVI and PAR, ( \alpha ) and ( \beta ) are parameter.</td>
</tr>
<tr>
<td>B</td>
<td>( GPP = \frac{(\gamma + NDVI) \times P_{\text{max}} \times \alpha \times PAR}{P_{\text{max}} + \alpha \times PAR} )</td>
<td>Modification of model A with ( P_{\text{max}} ) and NDVI as input variables, ( \gamma ) is a parameter. In another test, NDVI was replaced by WT or VWC as input variable.</td>
</tr>
<tr>
<td>C</td>
<td>( GPP = \frac{P_{\text{max}} \times \alpha \times PAR}{(P_{\text{max}})^2 \times (\alpha \times PAR)^2} )</td>
<td>Sigmoidal model after Wohlfahrt \textit{et al.} (2010).</td>
</tr>
<tr>
<td>D</td>
<td>( GPP = \frac{P_{\text{max}} \times PAR}{(\varepsilon + PAR)} \times T_2 \times (\sigma + LAI) )</td>
<td>Light response model (Marushchak \textit{et al.}, 2013); ( T_2 ) is the soil temperature at 2 cm soil depth and LAI is the leaf area index; ( \varepsilon ) and ( \sigma ) are parameter.</td>
</tr>
<tr>
<td>E</td>
<td>( R_{\text{eco,H}} = \frac{R_{\text{max}}}{1 + (\delta \times \exp^{-\kappa \times T})} )</td>
<td>Simple logistic function (Görrres \textit{et al.}, 2014, Rodeghiero &amp; Cescatti, 2005); ( R_{\text{max}} ) is the maximum respiration rate, ( \delta ) determines the elongation along the x-axis and ( \kappa ) affects the steepness of the curve at its inflection.</td>
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</table>

The GPP model was calibrated with PAR values measured inside the chamber. Although the transmissivity of the Plexiglas is high (> 92%) it causes a reduction of the radiation. During the complete measurement period, the PAR values inside the chamber (mean of the first ten seconds of the measurement) were on average 20% lower than the PAR values measured every minute outside the chamber. A lower PAR is causing a lower photosynthetic uptake inside the chamber. Therefore, modelling of GPP was conducted in two steps. At first, the GPP model was calibrated using PAR values measured inside the chamber, and secondly, the reproduction of GPP fluxes over the growing season was carried out using PAR values measured outside the chamber. Without this two-step calibration the CO\textsubscript{2} uptake rates would have been seriously underestimated. The NEE and R\textsubscript{A} fluxes were calculated as the sum of the modeled GPP and R\textsubscript{eco} fluxes and of the modeled R\textsubscript{eco} and R\textsubscript{H} fluxes, respectively. Furthermore, NPP was calculated from the sum of R\textsubscript{A} and GPP fluxes.
It was shown that CO₂ fluxes calculated with linear regression models can be seriously biased (Kutzbach et al., 2007a), while non-linear regression models significantly improve flux calculations (Pihlatie et al., 2013). On the other hand, Koskinen et al. (2014) have shown that, at least for respiration measurements, a linear regression model produced similar results and is more robust compared to other regression models, especially if chamber closure times are kept short. However, because the change in CO₂ and CH₄ concentration in the chamber was approximated best with a linear regression model (determined by AICc and R²adj) it was decided to run the models tentatively with two datasets. At first, only fluxes calculated with the linear regression were used for fitting. In a second step the fluxes which were calculated with the regression type which provided the best fit, determined by AICc, were used for fitting. This procedure revealed that model performance was best when fluxes were used, which were calculated with the linear regression model. On the basis of these findings, it was decided to use only the linear regression model for the calculation of chamber fluxes. This is in good agreement with Görres et al. (2014) who have shown that in some cases a linear regression model can produce a better CO₂ flux estimate for a non-linear concentration-over-time curve than a non-linear regression model.

4.5.2. CH₄ model

Similar to the CO₂ flux modelling, the measured CH₄ fluxes in 2015 were used to calibrate different CH₄ model approaches and the resulting fitting parameters were used to reproduce the CH₄ fluxes over the complete measurement period (Table 3). The functional relationship between environmental parameters and CH₄ emissions are known to be non-linear (Olefeldt et al., 2013). Therefore, several exponential multiple regression models were used to correlate the CH₄ fluxes with WT, VWC, ALD, Pmax (obtained from the GPP model) as well as different soil temperatures (at 2, 5, 10 and 15 cm soil depth) and the Tsurf.

None of these model approaches explained the CH₄ flux variation acceptably as they produced unrealistic fitting parameter and/or CH₄ fluxes. Therefore, to gap-fill the CH₄ fluxes over the complete measurement period the fluxes were integrated over time using linear interpolation for the days between the measurements.
Table 3: Failed CH₄ models for the reproduction of CH₄ fluxes. All listed functions produced unrealistic fitting parameter and/or fluxes.

<table>
<thead>
<tr>
<th>ID</th>
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<th>Remarks</th>
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<tbody>
<tr>
<td>A</td>
<td>( CH_4 = \alpha \times \beta \left[ \frac{(T-T_{ref})}{10} \right] )</td>
<td>Modified after Wille et al. (2008); input variables are soil or surface temperature. ( T_{ref} ) is the mean value of this variable during the measurement period; ( \alpha ) and ( \beta ) are parameter.</td>
</tr>
<tr>
<td>B</td>
<td>( CH_4 = \alpha \times \beta \left[ \frac{(T-T_{ref})}{10} \right] \times \exp(y \times WT, VWC) \times \exp(\kappa \times ALD) )</td>
<td>Modification of model A; Input variables are WT or VWC and ALD, ( y ) and ( \kappa ) are parameter.</td>
</tr>
<tr>
<td>C</td>
<td>( CH_4 = \alpha \times \beta \left[ \frac{(T-T_{ref})}{10} \right] \times \exp(y \times WT_{full}) )</td>
<td>Modification of model A; ( WT_{full} ) is the thickness of the water-saturated soil horizon (( WT_{full} = ALD - WT )).</td>
</tr>
<tr>
<td>D</td>
<td>( CH_4 = \alpha \times \beta \left( \frac{T-10}{10} \right) \times \exp(y \times WT) )</td>
<td>Exponential model after Marushchak et al. (2016).</td>
</tr>
<tr>
<td>E</td>
<td>( CH_4 = \alpha \times \exp(\beta \times T_{10}) \times \exp(y \times WT, \bar{P}_{max}) )</td>
<td>Exponential multiple regression model (Saarnio et al., 1997); ( T_{10} ) is the soil temperature at 10 cm depth, ( \bar{P}_{max} ) is the maximum canopy photosynthetic potential obtained from the CO₂ model.</td>
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</table>

To investigate the environmental parameters controlling CH₄ emissions, a curve estimation with a linear and an exponential model was performed between CH₄ fluxes from 2015 and 2014 and a set of environmental variables (ALD, WT, surface temperature, as well as soil temperature at 2 and 15 cm) using SPSS software (IBM Corp, Released 2014. IBM SPSS Statistics for Windows, Version 23.0, NY: IBM Corp.).
4.6. CO₂ and CH₄ budgets

To estimate the CO₂ and CH₄ budgets of the complete measurement period in 2015, the polygonal tundra on Samoylov Island was classified by Muster et al. (2012) into dry tundra (65%), wet tundra (19%) and water bodies (16%) (thermokarst lakes (> 0.1 hectare) are not included in this classification). These amounts were used to calculate the contribution of the single microsites to the overall budget for the polygonal tundra on Samoylov Island. For an estimate of the CO₂ fluxes, cumulative fluxes obtained from the model results were used. The cumulative CH₄ fluxes were calculated from linear interpolated daily mean values.
5. Results

5.1. Environmental conditions

5.1.1. Meteorological conditions

During the measurement period from mid-July until the end of August in 2014, the total precipitation was 30 mm, with most of the rainfall in August (28 mm) (Figure 12). The ALD at the polygonal center increased from 37 to a maximum of 43 cm on 18 August, while at the rim the ALD increased from 20 to 32 cm. The soil temperature at the polygonal rim had a higher diurnal amplitude than at the center and maximum soil temperatures were 28 °C and 17 °C at rim and center, respectively. The averaged air temperature during August was almost 11 °C, which is 2 °C higher as the long-term mean temperature for August (Boike et al., 2013). At the end of July, the air temperatures were highest with up to 30 °C. The WT ranged between -1 and -10 cm below the soil surface. Both WT and VWC fluctuations were tightly coupled with precipitation events (Figure 13). PAR reached continuously values of more than 1000 µmol m⁻² s⁻¹ during the measurement period (Figure 14). During the first week of the measurement period and at the beginning of August, the values were rather low for a few days.
5. Results

Figure 12: Soil temperatures, precipitation and active layer depths (ALD) at the study site in 2014. The upper graph shows the soil temperature at the polygonal center and rim as well as the daily precipitation rates during the measurement period. The bottom graph shows the evolution of ALD at both microsites.

Figure 13: Air and surface temperature as well as water table (WT) and volumetric water content (VWC) at the study site in 2014. The upper graph shows the air temperature measured at the EC system at 2 m height and the surface temperature calculated with equation (4). The bottom graph shows the VWC at the polygonal rim measured at 5 cm soil depth and the WT in relation to the soil surface at the polygonal center during the measurement period.
5. Results

**Figure 14**: Photosynthetic active radiation (PAR) at the study site during the measurement period in 2014. PAR covers the wavelength range between 400 and 700 nanometer and was measured in 30-min intervals at the eddy covariance (EC) system.

During the measurement period from mid-July to end of September in 2015, the soil temperatures at 2 cm depth at the polygonal rim showed a higher diurnal variability than at the center (Figure 15). Highest soil temperatures of almost 25 °C occurred in mid-July and at the beginning of August. During the same periods, the soil temperature at the center reached its maximum of just 15 °C. At the end of the measurement period, the temperatures at both microsites were slightly negative. The air temperature ranged between 23 °C and -2 °C (Figure 16). There were two warm periods with temperatures above 20 °C, one in mid-July and one at the beginning of August recorded. After both periods, the temperatures remained around 10 °C for more than one week. A third warm period was recorded from 5 to 7 September with temperatures of up to 20 °C. After this period temperature started to decrease continuously and reached a daily minimum of -2 °C on 23 September. The average temperature in August 2015 (9 °C) was similar to the long-term mean temperature between 1998-2011 (Boike et al., 2013), but about 2 °C colder than in 2014. Compared to the long-term mean, it was about 1°C colder during July (9 °C) but about 2 °C warmer during September (3 °C). A storm event with wind speeds of up to 11 m s⁻¹ and precipitation of 18 mm day⁻¹ reached the island around the 23 August (Figure 15). The total precipitation of the complete measurement period was 78 mm and at the end of September the precipitation was recorded as snow. While in 2014 28 mm of
5. Results

Rainfall was recorded between 3 and 20 August, the precipitation in 2015 during this period was almost 3-fold lower (10 mm). During July, nearly 50% (36 mm) of the total precipitation of the complete measurement period were recorded, while the total precipitation in September was just 9 mm. Longer periods without or with just light precipitation (≤ 0.1 mm day\(^{-1}\)) were recorded from 9 to 22 August and from 2 to 13 September.

ALD increased from the beginning of the campaign in mid-July until mid-September and then decreased slightly until the end of September (Figure 15). The ALD increased from 17 and 29 cm to a maximum of 36 and 37 cm at rim and center microsite, respectively. Maximum ALD was reached on 16 September at both microsites. At the polygonal center, the ALD nearly reached its maximum of 37 cm in July with depths of more than 35 cm and was almost constant at this depth until the values were decreasing at the end of September. In contrast to this, at the polygonal rim, the ALD was increasing almost constantly towards its maximum in September. During colder phases at the end of July and mid-August, the active layer depths were decreasing at both microsites. After the maximum was reached, the ALD decreased subsequently until the end of September by 3 cm and 4 cm at the polygonal center and the rim, respectively. WT fluctuations were coupled to precipitation. Throughout July, the WT was moving within few centimeters above the soil surface with a maximum of 5 cm on 21 July (Figure 16). During August, the water table dropped below the soil surface, but sharply increased after heavy rainfall on 23 August. Afterwards, accompanied by low precipitation, the WT decreased onto a minimum of 8 cm below the soil surface at the end of September. The volumetric water content at 5 cm soil depth was on average 30% at the polygonal rim. These values were peaking simultaneously to precipitation events.
5. Results

Figure 15: Soil temperatures, precipitation and active layer depths (ALD) at the study site in 2015. The upper graph shows the soil temperature at the polygonal center and rim as well as the daily precipitation rates during the measurement period. The bottom graph shows the evolution of ALD at both microsites.

Figure 16: Air and surface temperature as well as water table (WT) and volumetric water content (VWC) at the study site in 2015. The upper graph shows the air temperature measured at the EC system at 2 m height and the surface temperature calculated with equation (4). The bottom graph shows the VWC at the polygonal rim measured at 5 cm soil depth and the WT in relation to the soil surface at the polygonal center during the measurement period.

The photosynthetic active radiation showed a strong diurnal variation with lowest PAR values during night times (Figure 17). From the beginning of the campaign until 12 August, the mean values during nighttime (9 pm – 3 am) did not dropped under 5 \( \mu \text{mol m}^{-2} \text{s}^{-1} \).
Afterwards, with upcoming polar night conditions, the periods with low PAR values extended. High PAR values (PAR $> 1000$ $\mu$mol m$^{-2}$ s$^{-1}$) were measured in mid-July, at the beginning of August and once at the end of August. Throughout September, the daily maximum PAR values were decreasing. Similar to the temperatures at the end of July and mid-August, the PAR values were comparatively low during these periods.

**Figure 17:** Photosynthetic active radiation (PAR) at the study site during the measurement period in 2015. PAR covers the wavelength range between 400 and 700 nanometer and was measured in 30-min intervals at the eddy covariance (EC) system.
5. Results

5.1.2. Soil characteristics

At the polygonal rim, the soil bulk density increased sharply down the soil profile from 0.32 ± 0.03 g cm⁻³ in the organic-rich soil layer (0 to 6 cm) to 1.32 ± 0.08 g cm⁻³ in the mineral soil layer, while at polygonal center (0.36 ± 0.03 to 0.64 ± 0.01 g cm⁻³) this increase was not as sharp (Figure 18 and Figure 19). The gravimetric water contents were distinctly higher at the center than on the rim with 75 to 85% and 38 to 53%, respectively. The total C contents were lower at the polygonal rim (2-12%) compared to the center (10-20%) and decreased down the soil profile. Total SOC contents were decreasing distinctly with increasing depths at both microsites. The nitrogen content at the center was almost constant down the soil profile with around 0.6%. This constant nitrogen content leads together with the SOC content to a decreasing carbon nitrogen ratio from 33.1 to 16.9. In contrast to that, at the polygonal rim the nitrogen content was considerably higher in the organic-rich layer compared to the mineral soil layer (0.5% vs. 0.1%) and the difference of the C/N ratios between different soil depths is smaller, compared to the center. Strongly acidic pH values were measured at every measurement depth at the polygonal center (around 5.3), while in the organic-rich layer and in the mineral soil layer at the rim pH values were moderately acidic with values of 5.7 and 6.0, respectively.
### Polygongal center

**Location:** Samoylov Island, Lena River Delta  
**Geographic coordinates:** 72°22.442 N, 126°29.828 E  
**Field location:** center of low-center polygon  
**Date of profile acquisition:** 10.07.2014  
**Thaw depth during sampling:** 31 cm  
**Water level during sampling:** -2 cm  
**Vegetation:** sedge-moss tundra

**Remarks:** organic-layer very weakly decomposed, no evidence of cryoturbation

**Classification:**  
Russian classification (Elovskaya, 1987): Permafrost tundra humic-peatish

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<tr>
<th>Horizon donation</th>
<th>Depth</th>
<th>TC</th>
<th>TIC</th>
<th>TOC</th>
<th>TOC</th>
<th>N</th>
<th>C/N</th>
<th>pH</th>
<th>Water content</th>
<th>Dry bulk density</th>
<th>Further characteristics</th>
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<tr>
<td>Oi</td>
<td>0 to 6</td>
<td>19.8</td>
<td>0.2</td>
<td>19.6</td>
<td>4.2</td>
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<td>33.1</td>
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<td>0.2</td>
<td>13.0</td>
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<td>0.51</td>
<td>25.7</td>
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<td>75.4</td>
<td>0.53</td>
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<td>Ah2</td>
<td>15 to 23</td>
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<td>0.2</td>
<td>11.5</td>
<td>5.8</td>
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<td>77.1</td>
<td>0.62</td>
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<td>0.60</td>
<td>16.9</td>
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<td>0.64</td>
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<td>frozen</td>
</tr>
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</table>

![Soil profile diagram](image-url)
Figure 19: Soil characteristics and soil classifications of the soil from the polygonal rim.

**Polygonal rim**

**Location:** Samoylov Island, Lena River Delta

**Geographic coordinates:** 72°22.442 N, 126°29.828 E

**Field location:** elevated rim of low-center polygon

**Date of profile acquisition:** 10.07.2014

**Thaw depth during sampling:** 19 cm

**Water level during sampling:** -17 cm

**Vegetation:** moss-dominated tundra

**Remarks:** organic-rich layer weakly decomposed, evidence of cryoturbation (shown as @ in the horizon donation)

**Classification:**

*World Reference Base for Soil Resources (WRB, 2014):*

_**Turbic Glacial Cryosol** – abbreviation: CR-gl.tu_

_Russian classification (Elovskaya, 1987): Permafrost tundra silty-peatish_

<table>
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<th>Horizon donation</th>
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<th>TIC</th>
<th>TOC</th>
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<th>pH</th>
<th>Water content</th>
<th>Dry bulk density</th>
<th>Further characteristics</th>
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<td>Slightly decomposed plant material, some roots</td>
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<td>Ahl/@/Bf@</td>
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<td>6.7</td>
<td>0.2</td>
<td>6.5</td>
<td>2.4</td>
<td>0.27</td>
<td>23.9</td>
<td>5.97</td>
<td>48.2</td>
<td>0.92</td>
<td>Transitional horizon, humified organic matter, capillary fringe mottling, mineral fraction increases</td>
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<tr>
<td>Bf</td>
<td>10 to 19</td>
<td>2.1</td>
<td>0.2</td>
<td>1.9</td>
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<td>0.10</td>
<td>19.2</td>
<td>6.02</td>
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<td>1.32</td>
<td>Capillary fringe mottling, less organic matter, alpha-alpha-Dipyridyl negative, frozen</td>
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<td>19+</td>
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</tbody>
</table>
5. Results

5.1.3. Vegetation units

The wet polygonal center was dominated by the moss species *Drepanocladius revolvens* and *Meesia triqueta* as well as the sedge *Carex aquatilis* and marsh cinquefoil (*Comarum palustre*). In a few spots at the center, the sedge species *Carex chodorrhiza* and in the transition zone between polygonal center and rim partially the willow species *Salix glauca* occurred. The vegetation of the rim was made up mostly by the moss species *Hylocomium splendens* with a few parts consisting of *Polytrichum sp.* and *Rhytidium rugosum*. Vascular plants at this microsite were *Dryas punctata*, *Pyrola rotundifolia*, *Astragalus frigidus* and *Saussurea sp.* as well as some willow shrubs (*Salix glauca*, *Salix reticulata*). Furthermore, lichen species such as *Peltigera aptosa*, *Stereocaulon sp.* and *Cladonia rangiferina* were present at the drier spots (Table 4).
Table 4: Mapping of vegetation species and abundance at polygonal rim and center. The abbreviation R- and C- are representing rim and center plots, respectively. The analysis was done in four subplots (10 x 10 cm) of the measurement plots (50 x 50 cm). Abundances are given in % of the surface cover.

<table>
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<td></td>
<td>R-1</td>
<td>R-2</td>
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<tr>
<td><strong>Polygonal rim</strong></td>
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<tr>
<td><strong>Vascular plants</strong></td>
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<tr>
<td>Astragalus frigidus</td>
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<td>1</td>
</tr>
<tr>
<td>Pyrola rotundifolia</td>
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<td>12</td>
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<td>Salix reticulata</td>
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<td>6</td>
</tr>
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<td>Saxifraga cernua</td>
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<td>2</td>
</tr>
<tr>
<td>Dryas punctata</td>
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<td>1</td>
</tr>
<tr>
<td>Carex aquatilis</td>
<td>2</td>
<td>8</td>
</tr>
<tr>
<td>Saussurea sp.</td>
<td>4</td>
<td>4</td>
</tr>
<tr>
<td>Salix glauca</td>
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<td>-</td>
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<tr>
<td><strong>Lichens</strong></td>
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<tr>
<td>Cladonia rangiferina</td>
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<td>4</td>
</tr>
<tr>
<td>Peltigera aphhta</td>
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<tr>
<td>Stereocaulon sp.</td>
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<tr>
<td>Dactylina arctica</td>
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<td>-</td>
</tr>
<tr>
<td>Thamnolia vermicularis</td>
<td>&gt;95</td>
<td>&gt;95</td>
</tr>
<tr>
<td><strong>Moss</strong> (Hylocomium splendens, Polytrichum sp., Rhytidium rugosum)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>C-1</td>
<td>C-2</td>
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<tr>
<td><strong>Polygonal center</strong></td>
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<tr>
<td><strong>Vascular plants</strong></td>
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<tr>
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<td>Potentilla palustris</td>
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<tr>
<td>Salix glauca</td>
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<td>-</td>
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<tr>
<td>Carex chodorrhiza</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td><strong>Moss</strong> (Drepanocladus revolvens, Meesia triqueta)</td>
<td>&gt;95</td>
<td>&gt;95</td>
</tr>
</tbody>
</table>
5. Results

5.1.4. Porewater conditions

The pH of the soil porewater at the water-saturated polygonal center showed a high variability with neutral pH values of 6.8 and strongly acidic pH values of 5.5 (Figure 20). At the beginning of the measurement period, neutral pH values were measured close to the soil surface at 5 cm soil depth, while towards the end of the growing season, most neutral pH values were measured at the bottom of the active layer in 35 cm soil depth.

![Figure 20](image_url)

**Figure 20**: pH values of the soil porewater at the polygonal center in 2015. The pH was measured at different depths with an almost weekly interval. pH values from 35 cm soil depth were only measureable four times, as the soil was frozen at this depth until the end of August.
5. Results

5.2. CO₂ fluxes

5.2.1. Chamber fluxes 2014

The CO₂ fluxes measured in 2014 at the polygonal rim and center are shown in Figure 21. During the measurement period in 2014, the center acted as a robust net CO₂ sink with an averaged NEE of -54 ± 16 µg m⁻² s⁻¹. Highest net uptake of -83 ± 25 µg m⁻² s⁻¹ were measured at 3 August, while on 19 August a NEE of -25 ± 12 µg m⁻² s⁻¹ displayed the lowest uptake. In contrast, the polygonal rim was just a slight net CO₂ sink with an average of -2 ± 8 µg m⁻² s⁻¹. On 3, 13 and 19 August the rim was a net source for atmospheric CO₂ with a NEE of more than 20 µg m⁻² s⁻¹, while the highest net CO₂ uptake of -16 ± 5 µg m⁻² s⁻¹ was measured on 20 August. The daily averaged $R_{eco}$ fluxes were higher at the polygonal center (69 ± 22 µg m⁻² s⁻¹) compared to the rim (45 ± 4 µg m⁻² s⁻¹). They ranged between 31 ± 4 and 116 ± 31 µg m⁻² s⁻¹ and 25 ± 6 and 73 ± 7 µg m⁻² s⁻¹ at the center and rim, respectively. At the end of July, smoke of a forest fire was blown towards Samoylov Island for about one week. This likely affected GPP during this time as NEE was low even though PAR and temperatures (Figure 13 + Figure 14) were favorable for high CO₂ uptake rates. The calculated mean GPP fluxes were -48 ± 20 µg m⁻² s⁻¹ and -123 ± 24 µg m⁻² s⁻¹ at the rim and center, respectively. Highest GPP fluxes of -162 ± 27 µg m⁻² s⁻¹ were measured at the center on 3 August and on 9 August at the rim (-74 ± 14 µg m⁻² s⁻¹). The lowest GPP fluxes were measured at the rim on 13 August with just -4 ± 33 µg m⁻² s⁻¹. At the polygonal center, the lowest GPP fluxes of -80 ± 9 µg m⁻² s⁻¹ were measured at the beginning of the measurement period.

The mean $R_h$ fluxes were 30 ± 12 µg m⁻² s⁻¹ and 37 ± 14 µg m⁻² s⁻¹ at the polygonal center and rim, respectively. The lowest $R_h$ fluxes were measured at the beginning of the measurement period at both microsites, while the highest $R_h$ fluxes were measured simultaneously to highest air temperatures at the end of July. The averaged contribution of $R_h$ to $R_{eco}$ was 42% at the center. A distinctly higher contribution was observed at the rim with 85% and at more than half of the measurement days the measured $R_h$ fluxes exceeded the $R_{eco}$ fluxes. Based on these findings, $R_A$ and NPP fluxes were not calculated from the sum of $R_{eco}$ and $R_h$ fluxes as well as GPP and $R_A$ fluxes.
Figure 21: Chamber measurements of NEE, $R_{eco}$ and $R_{hi}$ in 2014. The colored error bars denote the standard deviation between the four replicate measurements. The standard error of the replicate measurements is on average 3.6 µg m$^{-2}$ s$^{-1}$ at the center and 2.1 µg m$^{-2}$ s$^{-1}$ at the rim. Both NEE ($n = 12$) and $R_{eco}$ fluxes ($n = 12$) were higher (in case of NEE more negative) at the center (right), while $R_{hi}$ fluxes ($n = 9$) were similar at both microsites.

5.2.2. Chamber fluxes 2015

The measured CO$_2$ fluxes in 2015 at the center and the rim are shown in Figure 22 and Figure 23, respectively. The standard error of the flux calculation was about 3.5 and 2.3 µg m$^{-2}$ s$^{-1}$ for polygonal center and rim, respectively and decreased slightly towards the end of the season. The net CO$_2$ uptake at the center was generally higher than at the rim (Figure 22). Highest NEE fluxes were measured at the end of July with $-97.1 \pm 27.0$ µg m$^{-2}$ s$^{-1}$ and $-208.6 \pm 17.0$ µg m$^{-2}$ s$^{-1}$ at the rim and center, respectively. In September, both microsites turned into small net sources for atmospheric CO$_2$. The highest net CO$_2$ release at the polygonal rim was measured on 17 August with $15.7 \pm 4.7$ µg m$^{-2}$ s$^{-1}$ and at the polygonal center on 19 September with $22.3 \pm 3.2$ µg m$^{-2}$ s$^{-1}$.

In contrast to the NEE fluxes, the measured $R_{eco}$ fluxes were on average higher at the rim compared to the center. Lowest ecosystem respiration fluxes at the polygonal center were observed on 23 July with $10.0 \pm 2.5$ µg m$^{-2}$ s$^{-1}$ and at the polygonal rim on 21 September.
5. Results

with 17.3 ± 1.3 µg m⁻² s⁻¹. Highest ecosystem respiration fluxes of 79.7 ± 10.8 and 88.1 ± 10.4 µg m⁻² s⁻¹ for rim and center, respectively, were measured at 9 August, when temperatures reached more than 20 °C.

NEE and $R_{eco}$ fluxes showed a clear seasonal trend. From mid-July the net CO₂ uptake increased until it peaked during the vegetation maximum at the end of July and beginning of August. Subsequently, NEE decreased until the end of September. This seasonality was more pronounced at the polygonal center than at the polygonal rim. Interestingly, the net CO₂ uptake at the polygonal rim was increasing from mid-September until the end of the measurement period. $R_{eco}$ showed almost the same, but less distinct seasonal pattern and the peak of the highest $R_{eco}$ fluxes was later in mid-August. In contrast, $R_{H}$ followed no seasonal trend.

Figure 22: Chamber measurements of NEE, $R_{eco}$ and $R_{H}$ fluxes at the polygonal center. The error bars denote the standard deviation of the four replicate measurements. The averaged standard deviation of the flux calculation during the measurement period was 3.5 µg m⁻² s⁻¹. Throughout the measurement period a pronounced seasonality of the NEE ($n = 83$) and $R_{eco}$ fluxes ($n = 85$) was observed, while $R_{H}$ fluxes ($n = 85$) showed no seasonal trend.
5. Results

Figure 23: Chamber measurements of NEE, $R_{\text{eco}}$ and $R_{\text{H}}$ fluxes at the polygonal rim. The error bars denote the standard deviation of the four replicate measurements. The averaged standard deviation of the flux calculation during the measurement period was $2.3 \, \mu\text{g} \, \text{m}^{-2} \, \text{s}^{-1}$. Similar to the polygonal center, the NEE ($n = 83$) and $R_{\text{eco}}$ fluxes ($n = 85$) at the rim showed a seasonality, while no seasonal trend was observed for the $R_{\text{H}}$ fluxes ($n = 85$).

The calculated GPP fluxes are naturally linked to determined fluxes of NEE and $R_{\text{eco}}$ as they are derived from those values (Figure 24 + Figure 25). Lowest GPP fluxes of $-10.4 \pm 2.5 \, \mu\text{g} \, \text{m}^{-2} \, \text{s}^{-1}$ and $-16.2 \pm 5.7 \, \mu\text{g} \, \text{m}^{-2} \, \text{s}^{-1}$ for center and rim microsite, respectively, were observed at the end of the campaign were senescence of plants is well advanced. Maximum GPP fluxes are associated with the mature state of plants at the end of July. The highest GPP flux at the polygonal rim was measured on 27 July with $-142.5 \pm 33.4 \, \mu\text{g} \, \text{m}^{-2} \, \text{s}^{-1}$, while at the polygonal center a maximum GPP flux of $-244.8 \pm 18.6 \, \mu\text{g} \, \text{m}^{-2} \, \text{s}^{-1}$ was measured on 2 August.

The calculated $R_{A}$ fluxes at the polygonal center were on average $18.2 \pm 13.9 \, \mu\text{g} \, \text{m}^{-2} \, \text{s}^{-1}$ with the highest fluxes of $56.0 \pm 10.0 \, \mu\text{g} \, \text{m}^{-2} \, \text{s}^{-1}$ measured on 17 August. Occasionally, the $R_{A}$ calculation revealed negative fluxes when $R_{H}$ fluxes were exceeding $R_{\text{eco}}$ fluxes, which occurred mostly during times of a high water table. At the polygonal rim, the averaged calculated $R_{A}$ flux was $15.6 \pm 8.2 \, \mu\text{g} \, \text{m}^{-2} \, \text{s}^{-1}$. At this microsite, the highest $R_{A}$ fluxes of
Results

$41.6 \pm 7.3 \mu g m^{-2} s^{-1}$ were observed on 18 July and the lowest $R_A$ fluxes were observed at mid-September with $3.1 \pm 8.6 \mu g m^{-2} s^{-1}$.

The calculated NPP fluxes showed, similar to the GPP fluxes, a distinct seasonality (Figure 24 and Figure 25). The mean NPP fluxes were $-94 \pm 61$ and $-55 \pm 26 \mu g m^{-2} s^{-1}$ at the polygonal center and rim, respectively. The highest incorporation of CO$_2$ into the soil-plant continuum at the center was on 30 July with $-222 \pm 18 \mu g m^{-2} s^{-1}$, which is three days earlier than the maximum GPP flux was determined. Similar to the highest GPP fluxes, the highest incorporation of CO$_2$ at the rim was determined on 27 July with $-115 \pm 29 \mu g m^{-2} s^{-1}$. The lowest NPP fluxes were determined in September with $-10 \pm 11 \mu g m^{-2} s^{-1}$ at the rim and $-2 \pm 6 \mu g m^{-2} s^{-1}$ at the center.

Figure 24: GPP, NPP and $R_A$ fluxes at the polygonal center in 2015. GPP fluxes ($n = 83$) were calculated from NEE-$R_{eco}$, $R_A$ fluxes ($n = 85$) were calculated from $R_{eco}$-$R_H$ and NPP fluxes ($n = 83$) were calculated from GPP-$R_A$. 

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Figure 25: GPP, NPP and $R_A$ fluxes at the polygonal rim in 2015. GPP fluxes ($n = 83$) were calculated from NEE-$R_{eco}$, $R_A$ fluxes ($n = 85$) were calculated from $R_{eco}$-$R_H$ and NPP fluxes ($n = 83$) were calculated from GPP-$R_A$.

Highest releases of CO$_2$ by $R_H$ were measured on 9 August at the polygonal center and rim with $37.6 \pm 5.5$ and $51.4 \pm 12.2$ µg m$^{-2}$ s$^{-1}$, respectively. The lowest $R_H$ fluxes were measured at both microsites in September. On 3 September remarkably low $R_H$ fluxes of $2.6 \pm 1.1$ µg m$^{-2}$ s$^{-1}$ were measured at the center, while at the rim lowest $R_H$ fluxes of $9.8 \pm 3.0$ µg m$^{-2}$ s$^{-1}$ were observed at the end of September. Increased $R_H$ fluxes after periodical re-clipping of the vegetation were not observed. The comparison of $R_H$ fluxes from measurement plots that were trenched in 2014 with those that were trenched in 2015 revealed no significant differences (t-test, $p > 0.05$) between the years of root-trenching (Figure 26).
Figure 26: Comparison of $R_H$ fluxes from plots trenched in 2014 and 2015. The measured fluxes are given with the standard deviation of the replicate measurements. The averaged standard error of the flux calculation is 1.0 and 1.5 $\mu$g m$^{-2}$ s$^{-1}$ at the center ($n = 73$) and rim ($n = 70$) that were trenched in 2015, respectively and 1.8 and 6.4 $\mu$g m$^{-2}$ s$^{-1}$ at the center ($n = 76$) and rim ($n = 76$) that were trenched in 2014, respectively. A student’s t-test revealed no significant difference ($p > 0.05$) between plots that were trenched in 2014 and 2015.

The mean contribution of $R_H$ to $R_{eco}$ over the complete measurement period calculated from the flux measurements was 44% at the polygonal center and 61% at the rim. Both values showed no seasonal trend. High contributions of $R_H$ to total ecosystem respiration were correlated with high WT at the polygonal center (Figure 27). During periods of lowest WT, the $R_H/R_{eco}$ ratio increased again. Furthermore, $R_A$ fluxes were low simultaneously to a high WT, but increased with a lowering of the WT, and decreased again, when the WT was at its minimum. In contrast to this, neither a trend to higher $R_H$ fluxes during high water levels nor to lower $R_H$ fluxes during low water levels was observed.
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Figure 27: Relationship between water table fluctuations with the $R_H/R_{eco}$ ratio and the $R_A$ fluxes at the polygonal center. Negative values on the x-axis indicate a water table below the soil surface. The relationships are best described by a polynomial model with higher $R_H/R_{eco}$ ratios ($n = 85$) and lower $R_A$ fluxes ($n = 85$) towards higher water levels as well as lower water levels.

5.2.3. Model parametrization

Based on the modeled CO$_2$ fluxes and meteorological conditions, the measurement period in 2015 can be divided into (1) a peak season (11 July – 20 August), defined by mature state of vascular plants, highest CO$_2$ fluxes and temperatures and (2) a post season (21 August – 23 September) defined by upcoming senescence of vascular plants as well as lower temperatures and CO$_2$ fluxes.

The fitting parameters $\alpha$ and $P_{max}$ as well as $R_{base}$ showed a strong spatial and temporal variability (Figure 28) during the measurement period. In general, $R_{base}$ were higher if the $R_{eco}$ and $R_H$ model was used to reproduce fluxes at the rim. The averaged $R_{base}$ values of the $R_H$ model did not differ significantly between peak and post season for center ($15.4 \pm 2.0 \mu g m^{-2} s^{-1}$ vs. $13.7 \pm 1.8 \mu g m^{-2} s^{-1}$) and rim ($28.7 \pm 3.7 \mu g m^{-2} s^{-1}$ vs. $29.3 \pm 1.6 \mu g m^{-2} s^{-1}$), but differed significantly between the microsites. Similarly, the averaged $R_{base}$ values of the $R_{eco}$ model differed between polygonal rim and center, but not significantly between the seasons (peak season: $18.7 \pm 2.0 \mu g m^{-2} s^{-1}$ vs.
5. Results

57.2 ± 7.2 µg m\(^{-2}\) s\(^{-1}\); post season: 15.5 ± 2.6 µg m\(^{-2}\) s\(^{-1}\) vs. 56.1 ± 7.5 µg m\(^{-2}\) s\(^{-1}\), for center and rim, respectively). For the GPP model, α values (mean: 1.47 ± 0.62) showed a high temporal variability with an average of 1.99 ± 0.18 during the peak season and 0.88 ± 0.34 throughout the post season, when plant senescence occurred. The \(P_{\text{max}}\) values showed a strong temporal variability when the GPP fluxes at the center were reproduced. Here, the mean \(P_{\text{max}}\) for the peak season was 334.4 ± 58.3 µg m\(^{-2}\) s\(^{-1}\), while during the post season the averaged \(P_{\text{max}}\) was 156.0 ± 35.1 µg m\(^{-2}\) s\(^{-1}\). Considerable differences in \(P_{\text{max}}\) were also observed between rim and center. The averaged values at the rim for the peak season was with 150.9 ± 25.3 µg m\(^{-2}\) s\(^{-1}\) much lower compared to the center. During the post season this difference was not as high as during the peak season; the averaged \(P_{\text{max}}\) at the rim was 119.0 ± 41.7 µg m\(^{-2}\) s\(^{-1}\). Similar to the measured NEE fluxes, \(P_{\text{max}}\) was increasing at the rim towards the end of the measurement period and the highest \(P_{\text{max}}\) of 203.9 ± 151.3 µg m\(^{-2}\) s\(^{-1}\) was observed at the last day of the measurement period.

**Figure 28:** Fitting parameters of the CO\(_2\) models. The values are given with the standard deviation of the model results of the single measurement plots (light grey error bars) and the confidence intervals (95%) of the fitting parameters (dark grey error bars). Panel (a) shows the initial canopy quantum efficiency α. The \(P_{\text{max}}\) values (panel b) show large differences between the polygonal center (white squares) and the rim (black triangles). Panel (c) shows \(R_{\text{base}}\) from the \(R_{\text{eco}}\) model and panel (d) \(R_{\text{base}}\) from the \(R_{\text{Hi}}\) model. Both models show higher \(R_{\text{base}}\) values for the rim.

For both microsites the \(R_{\text{eco}}\) model with an additional parameter (WT or VWC) suited best with a qualifying parameter \(R^2_{\text{adj}}\) of 0.46 and 0.78 for rim and center microsite, respectively.
respectively. Furthermore, the fluxes from the polygonal center (Figure 29) were best described by surface temperature, while for the polygonal rim (Figure 30) the soil temperature showed the best fitting compared to $T_{\text{soil}}$ and $T_{\text{surf}}$. In contrast to the $R_{\text{eco}}$ model the $R_{\text{H}}$ model without an additional parameter produced the highest $R^2_{\text{adj}}$ of 0.45 and 0.55 for polygonal rim and center, respectively, but the differences between the models in $R^2_{\text{adj}}$ were small. The fluxes from the center were best described by air temperature, while at the rim the soil temperature was the parameter with the highest explanatory power. The $R^2_{\text{adj}}$ of the GPP model was 0.82 for the polygonal center and 0.45 for the polygonal rim.

5.2.4. Modeled CO$_2$ fluxes

The modeled $R_{\text{eco}}$ ranged between 12 ± 3 µg m$^{-2}$ s$^{-1}$ and 69 ± 7 µg m$^{-2}$ s$^{-1}$ with an average of 30 ± 4 µg m$^{-2}$ s$^{-1}$ at the polygonal center during the peak season (Figure 29). During the post season, the averaged $R_{\text{eco}}$ values dropped slightly to 29 ± 3 µg m$^{-2}$ s$^{-1}$ with a range between 15 ± 1 µg m$^{-2}$ s$^{-1}$ and 52 ± 5 µg m$^{-2}$ s$^{-1}$. At the polygonal rim, the modeled $R_{\text{eco}}$ values were higher compared to the center throughout the first half of the growing season; they ranged between 23 ± 4 µg m$^{-2}$ s$^{-1}$ and 77 ± 14 µg m$^{-2}$ s$^{-1}$ with an average of 38 ± 7 µg m$^{-2}$ s$^{-1}$ (Figure 30). Afterwards, $R_{\text{eco}}$ decreased to an average of 32 ± 6 µg m$^{-2}$ s$^{-1}$ (21 ± 3 µg m$^{-2}$ s$^{-1}$ to 63 ± 11 µg m$^{-2}$ s$^{-1}$). The differences between the microsites were for both seasons not statistically significant (t-test, $p > 0.05$). Highest $R_{\text{eco}}$ values of the measurement period where encountered at both microsites at the beginning of August during a hot period with surface temperatures of more than 20 °C. High fluxes where also observed during the other hot periods (see section 5.1.1.). At the polygonal rim, the lowest $R_{\text{eco}}$ fluxes where obtained at the end of September accompanied by low soil temperatures. The lowest $R_{\text{eco}}$ values at the polygonal center where encountered on 21 July, associated with the highest water table during the campaign. A steep decrease in the modeled $R_{\text{eco}}$ fluxes at the center microsite was observed after the storm event on 23 August. In general, the values showed a larger diurnal amplitude at the rim compared to the center, which is in agreement with higher diurnal soil temperature variations at the rim (Figure 15).
At the polygonal rim, the modeled $R_H$ fluxes ranged between $14 \pm 4 \mu g \ m^{-2} \ s^{-1}$ and $46 \pm 13 \mu g \ m^{-2} \ s^{-1}$ during the peak season and $14 \pm 2 \mu g \ m^{-2} \ s^{-1}$ and $34 \pm 5 \mu g \ m^{-2} \ s^{-1}$ during the post season with averaged values of $22 \pm 5$ and $19 \pm 2 \mu g \ m^{-2} \ s^{-1}$, respectively (Figure 30). At the polygonal center, the $R_H$ fluxes were comparatively low; during the peak season they ranged between $8 \pm 1 \mu g \ m^{-2} \ s^{-1}$ and $27 \pm 2 \mu g \ m^{-2} \ s^{-1}$ with an average of $12 \pm 1 \mu g \ m^{-2} \ s^{-1}$ (Figure 29). Also throughout the post season, the modeled $R_H$ values were lower with $6 \pm 1 \mu g \ m^{-2} \ s^{-1}$ to $15 \pm 1 \mu g \ m^{-2} \ s^{-1}$ and an average of $9 \pm 1 \mu g \ m^{-2} \ s^{-1}$. The differences between the microsites were highly significant (t-test, $p < 0.001$) during the post season and significant ($p < 0.01$) during the peak season. The lowest $R_H$ fluxes of the measurement period at the center were encountered on 3 September accompanied by low air temperature. The highest value was observed during the first hot period in July. The highest and lowest $R_H$ fluxes at the polygonal rim were encountered at the same time as modeled $R_{eco}$ fluxes (8 August and 20 September). Similar to the $R_{eco}$ fluxes the $R_H$ fluxes at the rim had a larger diurnal amplitude compared to the center. By comparing the $R_H$ and the $R_{eco}$ model over the complete measurement period, the contribution of $R_H$ to $R_{eco}$ is on average 42% at the polygonal center and 60% at the rim. This contribution was distinctly higher during the peak season (47%) than throughout the post season (35%) at the center. At the polygonal rim, an opposed trend was observed, although the differences were with 57% during the peak season and 62% during the post season was substantial lower.

The modeled GPP fluxes showed a distinct seasonal trend. From the mid of July until 12 August gross primary productivity took place for 24 hours per day because of polar day conditions, even though the CO$_2$ uptake was low during night times. Afterwards, periods where gross primary productivity was zero extended due to arising polar night conditions. The averaged GPP fluxes for the peak season were $-138 \pm 11 \mu g \ m^{-2} \ s^{-1}$ and $-81 \pm 25 \mu g \ m^{-2} \ s^{-1}$ at the center and rim, respectively. At both microsites, the diurnal amplitude of GPP increased from the beginning of the campaign until it reached a maximum of $-163 \pm 57 \mu g \ m^{-2} \ s^{-1}$ and $-342 \pm 53 \mu g \ m^{-2} \ s^{-1}$ at the polygonal rim and center, respectively in mid-August. After this peak, the GPP values decreased continuously onto the lowest daily maximum of $-62 \pm 18 \mu g \ m^{-2} \ s^{-1}$ at 21 September at the center. The highest GPP fluxes during the post season were $-184 \pm 30 \mu g \ m^{-2} \ s^{-1}$
and \(-123 \pm 29 \mu g \ m^{-2} \ s^{-1}\) at the polygonal center and rim, respectively. Interestingly, the lowest daily maximum of GPP (\(-53 \pm 5 \mu g \ m^{-2} \ s^{-1}\)) at the rim microsite was observed about a week earlier than at the center microsite. Later, the GPP at the polygonal rim increased again. The averaged GPP fluxes during the post season were distinctly lower than during the peak season with \(-53 \pm 9 \mu g \ m^{-2} \ s^{-1}\) and \(-39 \pm 7 \mu g \ m^{-2} \ s^{-1}\) at the center and rim, respectively. The differences between the microsites were statistically significant (\(p < 0.01\)) during the peak season, but not during the post season (\(p > 0.05\)).

The modeled GPP and \(R_{\text{eco}}\) fluxes were used to calculate NEE fluxes (Figure 29 + Figure 30). Throughout the peak season, the highest net \(CO_2\) uptake was \(-117 \pm 60 \mu g \ m^{-2} \ s^{-1}\) at the rim (23 July) and \(-288 \pm 53 \mu g \ m^{-2} \ s^{-1}\) at the center (16 August). Highest net \(CO_2\) release was measured in mid-August with \(49 \pm 10 \mu g \ m^{-2} \ s^{-1}\) and \(54 \pm 2 \mu g \ m^{-2} \ s^{-1}\) at the polygonal rim and center, respectively. The averaged NEE flux at the center during the peak season was \(-107 \pm 12 \mu g \ m^{-2} \ s^{-1}\). At the polygonal rim, the averaged NEE flux was much lower with \(-44 \pm 27 \mu g \ m^{-2} \ s^{-1}\). From 11 July until the 28 July, the NEE at the polygonal center was continuously negative, while first net \(CO_2\) release at the polygonal rim was encountered on 17 July. The diurnal amplitude of NEE oscillation was greatest between the end of July and mid-August. In the post season, the NEE at the center ranged between \(-166 \pm 30 \mu g \ m^{-2} \ s^{-1}\) and \(43 \pm 4 \mu g \ m^{-2} \ s^{-1}\). At the polygonal rim, the maximum net uptake was lower compared to the center with \(-92 \pm 46 \mu g \ m^{-2} \ s^{-1}\), while the highest net release was similar with \(43 \pm 8 \mu g \ m^{-2} \ s^{-1}\). Compared to the peak season, the averaged net \(CO_2\) uptake was much lower during the post season with values of \(-23 \pm 10 \mu g \ m^{-2} \ s^{-1}\) and \(-7 \pm 11 \mu g \ m^{-2} \ s^{-1}\) for polygonal center and rim, respectively. Highest net uptake during the post season were encountered at the end of August. From that period on the net \(CO_2\) uptake at the center decreased and at some days the NEE became positive on a daily average but even at the end of the measurement period a net \(CO_2\) uptake was observed during daytime. Similar to the GPP fluxes, the NEE fluxes at the rim were increasing again towards higher net \(CO_2\) uptake from mid-September until the end of the measurement period.

In the peak season, the \(R_\lambda\) fluxes were ranging between \(1 \pm 3 \mu g \ m^{-2} \ s^{-1}\) and \(55 \pm 4 \mu g \ m^{-2} \ s^{-1}\) at the polygonal center (Figure 29) and between \(5 \pm 5 \mu g \ m^{-2} \ s^{-1}\) and \(32 \pm 19 \mu g \ m^{-2} \ s^{-1}\) at the rim (Figure 30). On average, the \(R_\lambda\) fluxes were slightly higher at
the center with 18 ± 4 µg m\(^{-2}\) s\(^{-1}\) compared to 16 ± 9 µg m\(^{-2}\) s\(^{-1}\) at the rim. The lowest \(R_A\) fluxes were accompanied by a low VWC (rim) and a high water table (center). During the post season, the lowest \(R_A\) flux at the center was observed directly after the storm event with 5 ± 1 µg m\(^{-2}\) s\(^{-1}\) and at the rim at the end of the campaign (6 ± 4 µg m\(^{-2}\) s\(^{-1}\)). The maximum \(R_A\) flux in the post season of 40 ± 4 µg m\(^{-2}\) s\(^{-1}\) was observed at the center during the warm period in September. At the rim, the \(R_A\) flux was highest shortly after the storm event with 30 ± 12 µg m\(^{-2}\) s\(^{-1}\), accompanied by a high VWC. On average, the \(R_A\) fluxes were 19 ± 3 µg m\(^{-2}\) s\(^{-1}\) and 13 ± 6 µg m\(^{-2}\) s\(^{-1}\) during the post season at the polygonal center and rim, respectively.

The NPP fluxes were calculated from the sum of GPP and \(R_A\) fluxes. In general, the NPP fluxes are closely linked to variations of PAR values. The averaged NPP fluxes during the peak season were -122 ± 12 µg m\(^{-2}\) s\(^{-1}\) and -66 ± 28 µg m\(^{-2}\) s\(^{-1}\) at the center and rim, respectively. At the polygonal center, the highest NPP fluxes during the peak season were determined on 16 August with -300 ± 53 µg m\(^{-2}\) s\(^{-1}\) and throughout the post season on 24 August with -178 ± 30 µg m\(^{-2}\) s\(^{-1}\). The lowest daily maximum at the center was observed on 19 August with -114 ± 17 µg m\(^{-2}\) s\(^{-1}\) and in mid-September with -39 ± 14 µg m\(^{-2}\) s\(^{-1}\) during the peak and post season, respectively. At the rim on 13 July (-89 ± 40 µg m\(^{-2}\) s\(^{-1}\)) and on 14 September (-43 ± 8 µg m\(^{-2}\) s\(^{-1}\)) the lowest daily maxima were determined for peak and post season, respectively. The highest NPP fluxes at the rim during the peak season were observed on 16 August with -142 ± 57 µg m\(^{-2}\) s\(^{-1}\). During the post season, the maximum NPP fluxes were determined rather late in the season on 22 September with -107 ± 46 µg m\(^{-2}\) s\(^{-1}\).
Figure 29: Modeled and measured CO₂ fluxes at the polygonal center in 2015. Measured fluxes are available for NEE, $R_{eco}$ and $R_h$. NEE model fluxes are calculated from GPP-$R_{eco}$, $R_A$ model fluxes from $R_{eco}-R_h$ and NPP model fluxes from GPP-$R_A$. Note the different scales.
Figure 30: Modeled and measured CO₂ fluxes at the polygonal rim in 2015. Measured fluxes are available for NEE, $R_{eco}$ and $R_H$. NEE model fluxes are calculated from GPP-$R_{eco}$, $R_{A}$ model fluxes from $R_{eco}$-$R_H$ and NPP model fluxes from GPP-$R_{A}$. Note the different scales.
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5.2.5. Cumulative CO₂ fluxes

Based on the modeled CO₂ fluxes, the cumulative CO₂ fluxes were calculated (Table 5). The cumulative net CO₂ uptake during the peak season is more than twice as high at the center compared to the rim. During the post season, the differences in NEE fluxes between the microsites increased slightly. Partitioning of NEE fluxes into the underlying processes revealed that the cumulative GPP flux at the polygonal center is higher than at the polygonal rim during the peak season. Also throughout the post season, the cumulative GPP flux at the center is higher compared to the rim, but the gap between these fluxes decreased sharply (Table 5). Interestingly, the cumulative Reco fluxes at the rim are higher than at the center. Similar to Reco, the cumulative RH fluxes were higher at the rim compared to the center during peak and post season. The cumulative GPP and RA fluxes were used to calculate the net primary productivity (NPP). Over the complete measurement period, the NPP was almost twice as high at the center compared to the rim, but similar to the GPP fluxes the gap in NPP between the microsites was distinctly lower in the post season. At the center, the cumulative NPP was more than four-fold higher during the peak season compared to the post season. Similar to the center, more carbon was accumulated in the soil at the rim throughout the peak season compared to the post season.

Table 5: Cumulative CO₂ fluxes in 2015. The values were calculated from the model results. The cumulative fluxes show distinct differences between the peak and the post season. The values are given in g C m⁻². In total, both microsites are representing a net CO₂ sink over the measurement period with highest NEE throughout the peak season, whereas during the post season the respiration fluxes gained more importance.

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5. Results

5.3. CH₄ fluxes

5.3.1. Chamber fluxes 2014

The measured CH₄ emissions in 2014 showed distinct differences between the microsites (Figure 31). The CH₄ fluxes at the center were with a mean of $0.45 \pm 0.03 \, \mu g \, m^{-2} \, s^{-1}$ more than one order of magnitude higher than the CH₄ emissions from the rim ($0.019 \pm 0.008 \, \mu g \, m^{-2} \, s^{-1}$). Highest CH₄ emissions at the rim were measured on 19 August with $0.045 \pm 0.005 \, \mu g \, m^{-2} \, s^{-1}$, while the lowest CH₄ fluxes at this microsite of $0.01 \pm 0.004 \, \mu g \, m^{-2} \, s^{-1}$ were measured just one day after the maximum on 20 August. The CH₄ emissions at the center showed a much lower range over the measurement period compared to the rim. The lowest emissions of $0.36 \, \mu g \, m^{-2} \, s^{-1}$ were measured on 9 August and highest emissions at the same day a few hours later ($0.51 \pm 0.05 \, \mu g \, m^{-2} \, s^{-1}$).

![Figure 31: Measured CH₄ fluxes at polygonal rim and center in 2014. The error bar denote the standard deviation between the four replicate measurements. The averaged standard error of the flux calculation was $0.026$ and $0.002 \, \mu g \, m^{-2} \, s^{-1}$ at the center and rim, respectively. The measured CH₄ emissions were about one order of magnitude higher at the center (n = 34) compared to the rim (n = 34).]
5. Results

5.3.2. Chamber fluxes 2015

The results of the CH$_4$ flux chamber measurements in 2015 for the polygonal center and rim are shown in Figure 32. The standard error of the flux calculation was about 0.016 µg m$^{-2}$ s$^{-1}$ at the center and 0.002 µg m$^{-2}$ s$^{-1}$ at the rim. These values were slightly increasing towards the end of the measurement period. Both investigated microsites were a source for atmospheric CH$_4$, but the source strengths differ considerably.

![Figure 32: Measured CH$_4$ fluxes at the polygonal rim and center in 2015. The error bar denote the standard deviation between the four replicate measurements. The averaged standard error of the flux calculation was 0.016 and 0.002 µg m$^{-2}$ s$^{-1}$ at rim (n = 84) and center (n = 85), respectively. The fluxes show a pronounced seasonality at the center with highest fluxes at the beginning of September. With a mean of 0.30 ± 0.08 µg m$^{-2}$ s$^{-1}$ compared to 0.017 ± 0.004 µg m$^{-2}$ s$^{-1}$ the CH$_4$ emissions were more than one order of magnitude higher at the polygonal center than at the rim, respectively. The highest CH$_4$ emissions of 0.41 ± 0.024 µg m$^{-2}$ s$^{-1}$ were measured at the center on 13 September while the lowest emissions were measured on 11 July with just 0.065 ± 0.004 µg m$^{-2}$ s$^{-1}$. The CH$_4$ fluxes at the center showed a distinct seasonality with increasing emissions from July on until they reached a peak in mid-September. Afterwards, the CH$_4$ emissions were decreasing but were still rather high compared to emissions from July. At the polygonal rim, the same seasonal curve was observed, but less](image-url)
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pronounced as at the center. Here, CH$_4$ emissions were also increasing from the beginning of the measurement period, but no decrease was observed towards the end of September. The highest CH$_4$ flux at the rim was measured in mid-September with $0.027 \pm 0.003$ $\mu$g m$^{-2}$ s$^{-1}$ while the lowest emission of $0.006 \pm 0.002$ $\mu$g m$^{-2}$ s$^{-1}$ was measured on 15 July. Similar to the CO$_2$ fluxes, the measurement period can be divided into a peak and a post season. At both microsites, higher CH$_4$ emissions were measured during the post season. At the polygonal rim, the averaged mean CH$_4$ fluxes were $0.017 \pm 0.004$ $\mu$g m$^{-2}$ s$^{-1}$ during the peak season and $0.022 \pm 0.003$ $\mu$g m$^{-2}$ s$^{-1}$ throughout the post season. At the polygonal center the CH$_4$ emissions were about one third lower during the peak season ($0.24 \pm 0.08$ $\mu$g m$^{-2}$ s$^{-1}$) compared to those of the post season ($0.36 \pm 0.04$ $\mu$g m$^{-2}$ s$^{-1}$).

The comparison between the measured CH$_4$ fluxes from 2014 and 2015 revealed similar emissions at the polygonal rim. At the center, the mean CH$_4$ emissions in 2014 were higher than the highest single CH$_4$ fluxes measured in 2015.
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5.3.3. Correlation with environmental parameters

Overall, the CH$_4$ fluxes from 2015 at the polygonal rim and center were poorly explained with environmental variables (Table 6). For both microsites, the environmental variable with the highest explanatory power for the CH$_4$ emissions was the ALD. However, the ALD explained just 27% and 45% of the CH$_4$ flux variance at the rim and center, respectively. Furthermore, no clear difference between the tested linear and exponential model was observed. Smaller explanatory power were given by the WT fluctuations and $T_{soil}$ at 2 cm for CH$_4$ emissions from the center. At the polygonal rim, no further substantial correlation (> 20%) with environmental variables was observed.

The CH$_4$ fluxes from 2014 at the polygonal rim and center did not correlate significantly with any of the environmental variables. Neither a linear regression nor an exponential regression model was statistically significant ($p < 0.05$).

Table 6: Linear and exponential relationship between CH$_4$ emissions and environmental parameters in 2015. All relationships at the center ($n = 331$) and rim ($n = 322$) were statistically significant (significance level < 0.001). The goodness of the relationships are described by the $R^2_{adj}$.

<table>
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<td>0.07</td>
<td>$T_{surf}$ exp</td>
<td>0.04</td>
</tr>
<tr>
<td>$T_{soil}$ 2 cm lin</td>
<td>0.20</td>
<td>$T_{soil}$ 2 cm lin</td>
<td>0.03</td>
</tr>
<tr>
<td>$T_{soil}$ 2 cm exp</td>
<td>0.23</td>
<td>$T_{soil}$ 2 cm exp</td>
<td>0.03</td>
</tr>
<tr>
<td>$T_{soil}$ 15 cm lin</td>
<td>0.03</td>
<td>$T_{soil}$ 15 cm lin</td>
<td>0.01</td>
</tr>
<tr>
<td>$T_{soil}$ 15 cm exp</td>
<td>0.04</td>
<td>$T_{soil}$ 15 cm exp</td>
<td>0.01</td>
</tr>
</tbody>
</table>
6. Discussion

This discussion focusses on CO₂ and CH₄ fluxes measured in 2015. The flux data from 2014 are just sporadically discussed. The dataset, the variability of fluxes and environmental parameters as well as the measurement period were too small and too short in 2014, which renders a profound interpretation of the data impossible.

6.1. CO₂ fluxes

In this study, the first values of net ecosystem exchange, gross primary productivity, net primary productivity as well as ecosystem respiration, heterotrophic respiration and autotrophic respiration fluxes obtained from modelling approaches for different microsites of the polygonal tundra were presented (Figure 29 and Figure 30). These fluxes are of crucial importance as they show the different response of the underlying processes of NEE to environmental controls over the growing season on a spatial and temporal scale. Good correlations were observed of GPP fluxes with PAR. Rₑₑₒ fluxes correlated with soil and surface temperature, but also with VWC and WT for rim and center, respectively. However, for Rₐ fluxes the correlations with environmental parameters were smaller than for GPP and Rₑₑₒ fluxes, which demonstrates the complexity and the still limited understanding of the factors controlling these fluxes. Both microsites acted as net sinks for atmospheric CO₂ during the measurement period in 2015, which supports Hypothesis 1. The CO₂ sink strength differed substantially between the microsites (Table 5). This difference is most likely mainly in response to different hydrological conditions, but also to soil temperature and vegetation composition. This finding supports Hypothesis 2, that contrasting hydrological conditions mainly drive the differences in CO₂ fluxes. The cumulative Rₑₑₒ fluxes at the rim are higher as at the center. This is remarkable as high Rₑₑₒ fluxes are generally associated with high GPP fluxes (Bubier et al., 2003), since CO₂ uptake via photosynthesis displays the source of Rₐ fluxes. This might be caused by comparatively low Rₐ fluxes at the polygonal center. Despite substantial higher GPP fluxes, the Rₐ fluxes at the center are within the same range with those from the rim. The comparatively low Rₐ fluxes at the center might be an effect of contrasting hydrological conditions between both microsites. In total, these differences lead to an almost two-times higher NPP at the polygonal center in relation to the rim.
6. Discussion

6.1.1. CO₂ fluxes in comparison with other studies

The daily averaged CO₂ fluxes obtained from the CO₂ models were compared with CO₂ fluxes from other arctic tundra sites that are similar in vegetation and soil composition to the polygonal tundra in the Lena River Delta (Table 7). It has to be noted that measurement periods, plant growth forms and density as well as climatic conditions can differ among the studies. These differences can lead to differences in CO₂ fluxes. However, all values are representing at least a rough estimate of growing season CO₂ surface-atmosphere fluxes of an arctic tundra ecosystem.

The $R_{eco}$ fluxes at both microsites are at the lower end in comparison with other arctic tundra sites (Table 7). Solely, a wet tundra site in the Komi Republic, Russia (Heikkinen et al., 2004), a wet sedge site at Daring Lake, Canada (Nobrega & Grogan, 2008) and a polygonal center site (Oechel et al., 1995) showed $R_{eco}$ fluxes that were within the same range as in this study. The low $R_{eco}$ fluxes reported from this study lead to relatively high NEE fluxes at the polygonal center, benefited by moderate GPP fluxes, compared to other tundra sites. Furthermore, the comparison revealed that comparatively high estimates of NEE were reported solely from wet and sedge-dominated sites (Table 7), but these sites can also act as net source for atmospheric CO₂ (Oechel et al., 1995). The GPP fluxes in some arctic regions are distinctly higher than in the Lena River Delta (Olivas et al., 2011, Ström et al., 2012, Vourlitis et al., 2000, Zamolodchikov et al., 2000). The low GPP fluxes at the polygonal rim are causing, despite low $R_{eco}$ fluxes, a comparatively low NEE at the rim.
Table 7: Comparison of daily averaged CO$_2$ fluxes from arctic tundra sites. The sites are similar in vegetation and soil composition. All listed fluxes were measured with the closed chamber technique. *: standard error estimated

<table>
<thead>
<tr>
<th>Location</th>
<th>Tundra type</th>
<th>Period</th>
<th>NEE (g C m$^{-2}$ d$^{-1}$)</th>
<th>GPP (g C m$^{-2}$ d$^{-1}$)</th>
<th>$R_{eco}$ (g C m$^{-2}$ d$^{-1}$)</th>
<th>Ref</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lena River Delta, RU (72°N, 127°E)</td>
<td>pol. rim pol. center</td>
<td>Jul-Sep 2015</td>
<td>-0.6 ± 0.4</td>
<td>-1.4 ± 0.4</td>
<td>0.8 ± 0.2</td>
<td>a</td>
</tr>
<tr>
<td>Lek Vorkuta, RU (67°N, 63°E)</td>
<td>shrub sedge bog</td>
<td>Jul-Aug 1996</td>
<td>-0.6 ± 0.3</td>
<td>-4.5 ± 0.4</td>
<td>3.9 ± 0.3</td>
<td>b</td>
</tr>
<tr>
<td>Prudhoe Bay, US (70°N, 149°W)</td>
<td>pol. tundra</td>
<td>Jun-Aug 1994</td>
<td>-0.6 ± 0.4</td>
<td>-5.2 ± 0.6</td>
<td>4.6 ± 0.3</td>
<td>c</td>
</tr>
<tr>
<td>Lek Vorkuta, RU (67°N, 63°E)</td>
<td>wet peaty tundra</td>
<td>Jun-Sep 2001</td>
<td>-1.1 ± 0.2</td>
<td>-1.9 ± 0.2*</td>
<td>0.9 ± 0.2*</td>
<td>d</td>
</tr>
<tr>
<td>Daring Lake, CA (65°N, 111°W)</td>
<td>dry heath mesic birch wet sedge</td>
<td>Jun-Sep 2004</td>
<td>-0.01 ± 0.1</td>
<td>-1.7 ± 0.3</td>
<td>1.8 ± 0.2</td>
<td>e</td>
</tr>
<tr>
<td>Barrow, US (71°N, 157°W)</td>
<td>pol. rim pol. center</td>
<td>Jun-Aug 2005</td>
<td>-0.1 ± 0.5</td>
<td>-3.7 ± 0.2</td>
<td>3.6 ± 0.3</td>
<td>f</td>
</tr>
<tr>
<td>Barrow, US (71°N, 157°W)</td>
<td>pol. rim pol. center</td>
<td>Jun-Aug 2006</td>
<td>-0.2 ± 0.2</td>
<td>-3.1 ± 0.3</td>
<td>2.9 ± 0.1</td>
<td>f</td>
</tr>
<tr>
<td>Barrow, US (71°N, 157°W)</td>
<td>pol. center</td>
<td>Jun-Aug 1992</td>
<td>0.04 ± 0.05</td>
<td>-0.8 ± 0.1</td>
<td>0.8 ± 0.1</td>
<td>g</td>
</tr>
<tr>
<td>Zackenberg, GL (74°N, 20°W)</td>
<td>sedge-dom. fen</td>
<td>Jun-Aug</td>
<td>-2.6 ± 0.3</td>
<td>-5.6 ± 0.4</td>
<td>3.0 ± 0.1</td>
<td>h</td>
</tr>
</tbody>
</table>

a: This study; b: Zamolodchikov et al. (2000); c: Vourlitis et al. (2000); d: Heikkinen et al. (2004); e: Nobrega and Grogan (2008); f: Olivas et al. (2011); g: Oechel et al. (1995); h: Ström et al. (2012)

6.1.2. Environmental controls on CO$_2$ fluxes

The polygonal tundra on Samoylov Island in the Lena River Delta has to be considered as an ecosystem with low GPP due to low vascular plant coverage with a maximum leaf coverage of 0.3 (Kutzbach et al., 2007b). Mosses, which have a high leaf coverage (> 0.9), are dominant at both microsites and have, similar to lichens, a much lower photosynthetic capacity than vascular plants (Brown et al., 1980). Furthermore, photosynthesis of vascular plants in arctic tundra ecosystems is restricted by a low nutrient availability. This is true for most tundra soils due to often water-logged and cold soil conditions (Oechel et al., 1998), as these conditions ensure low microbial decomposition rates (Hobbie et al., 2002), which in turn lead to a low supply of bioavailable nutrients (Beermann et al., 2015). However, following the explanation of Kutzbach et al. (2007b), the soils at the study site can be considered as extremely nutrient-poor as the study site is one of the coldest permafrost regions on the earth (Romanovsky et al., 2010), the parent material consists of nutrient-poor sands and the Holocene river terrace is due to its elevation not regularly
flooded, so that fresh nutrient input via water transport is lacking. Additionally, the net radiation at the study site from June to August is with a mean of 85 W m\(^{-2}\) (1999-2011) lower than those reported from most other arctic tundra sites (Boike et al., 2013). These factors might explain the comparatively low GPP fluxes at the polygonal rim and center at the study site compared to other arctic tundra sites.

Differences in GPP fluxes between the rim and center are also related to the vascular plant coverage. The polygonal center has a much higher abundance of sedges while the rim is moss-dominated (Table 4) and the sparsely spread vascular plants have shorter and fewer leaves. Therefore, the photosynthetic capacity of the center is higher, which results in higher GPP. Additionally, limited water availability due to the elevation of the polygonal rim allows moisture to run off and desiccate the moss layer at the rim, which then would lead to decreasing GPP fluxes (Olivas et al., 2011). On the other hand, Olivas et al. (2011) found GPP fluxes to be higher at the polygonal rim than at the polygonal center (Table 7). They related low GPP fluxes at the center to submersion of the moss layer and vascular plants at the center, which requires a constant WT above the soil surface. At the polygonal center from this study, the WT was frequently below the soil surface so that submersion is impeded partwise. This difference in GPP fluxes between both study sites reveals the importance of the water level and fluctuations throughout the season as they obviously can have an impact on GPP fluxes. However, the addition of WT and VWC values as parameter did not improve the GPP model (Table 2). An impact of hydrological conditions on GPP fluxes in this study might be low due to missing continuous plant submersion or might be masked due to contrasting PAR values and WT/VWC fluctuations.

Differences in NEE fluxes between the microsites can also be related to their different soil conditions. Cold and water-logged soil conditions, such as in polygonal depressions like the centers, inhibits decomposition and mineralization of SOM due to oxygen limitation which causes low microbial activity (Hobbie et al., 2002). On the other hand, the moisture run-off at the rim creates dry conditions at the rim, which increases soil oxygen availability and therefore enhances \(R_{\text{eco}}\) and \(R_{\text{H}}\) (Oechel et al., 1998). The higher diurnal amplitude of the soil temperature (Figure 15), a product of the thermic buffer function of the standing water at the center, leads to higher daily soil temperatures at the polygonal rim compared to the center. These higher temperatures are causing higher decomposition rates and
therefore higher $R_H$ fluxes in relation to sites with high water levels. Hence, low NEE estimates at the rim are occurring not only because of lower GPP but also due to higher $R_{eco}$ fluxes compared to the center. In general, respiration fluxes from the polygonal tundra of the Lena River Delta are expected to be low since the factors controlling $R_{eco}$ and $R_H$ are not favorable for a high respiration rates at this site (Gorham, 1991). This finding is in good agreement with Nobrega and Grogan (2008) who compared a wet sedge with a dry heath and a mesic birch site and found that the NEE at the wet sedge site is highest (Table 7). They concluded that SOC accumulation in wet-sedge tundra is highest because respiration is restricted due to the water-logged conditions.

Interestingly, measurements of CO$_2$ fluxes at the polygonal rim show an increase of NEE throughout September. This increase cannot be explained with higher PAR or temperature during this time of the season and at the polygonal center the net CO$_2$ uptake was continuously decreasing during September. Instead, the increase of net CO$_2$ uptake at the rim towards the end of measurement period can be related to the photosynthetic activity of mosses. Mosses can remain photosynthetic active for years (Collins & Oechel, 1974) and Kutzbach et al. (2007b) considered the September at the EC footprint area as period where C uptake occurs mostly due to moss photosynthesis. During this time of the growing season, mosses can still assimilate substantial amounts of CO$_2$ because they tend to reach light saturation at lower irradiance (Harley et al., 1989). The photosynthetic activity of mosses declines rapidly when they face desiccation, because they cannot control their tissue water content (Turetsky et al., 2012). It was also shown that mosses face light stress during times of high PAR (Murray et al., 1993). This light stress causes delayed senescence and more late-season photosynthesis (Zona et al., 2011). Therefore, the photosynthetic activity at the polygonal rim is expected to be low during warm and dry weather periods like at the beginning of September (Figure 15) and during times of high PAR. With continuous rainfall, dew formation and lower PAR in mid-September, the mosses resume metabolic active, which leads to increasing net CO$_2$ uptake at the rim.

6.1.3. Heterotrophic respiration fluxes in arctic tundra ecosystems

The comparison of $R_H$ fluxes in 2015 from sampling plots that were trenched in 2014 to those that were trenched in 2015 showed no significant differences (Figure 26). In
6. Discussion

In contrast to that, the contribution of $R_H$ to $R_{eco}$ fluxes at the rim in 2014 was remarkably higher as in 2015 and the measured $R_H$ fluxes partwise exceeded the measured $R_{eco}$ fluxes. It is assumed that these high $R_H$ fluxes are an artefact of the root-trenching, most likely due to the additional decomposition of residual roots which is one of the main drawbacks of this method (Figure 4) (Subke et al., 2006). However, Shurpali et al. (2008) suggested that this artefact contribute little to $R_H$ fluxes several months after the treatment. It is therefore assumed, that the root-trenching method produced reliable $R_H$ fluxes at the study site in 2015 as no evidence of a significant over- or underestimation was observed one year after the treatment.

To date, there are just a few estimates of growing season $R_H$ fluxes from arctic tundra ecosystems (Table 8). Differences in $R_H$ fluxes between the tundra sites may be caused by different time of waiting after the treatment. Nobrega and Grogan (2008) started the $R_H$ measurements one day after clipping, while measurements for this as well as for the study of Biasi et al. (2014) started about one year after the treatment. Therefore, even though the partitioning approach for seasonal estimates of $R_H$ fluxes is similar for all studies, a comparison has to be handled with caution. The few $R_H$ flux estimates from other arctic tundra sites are comparatively higher than the $R_H$ values from the Lena River Delta (Table 8). Considerable higher $R_H$ fluxes throughout the growing season were measured at a mesc birch and a dry heath site at Daring Lake in Canada (Nobrega & Grogan, 2008) and at a bare peat site in the subarctic tundra at Seida, Russia (Biasi et al., 2014). Both sites contained substantial higher amounts of SOC in the organic-rich layer, which most likely explain higher $R_H$ fluxes due to higher decomposition rates. $R_H$ fluxes within the same range as in this study were solely measured at a wet sedge site in Daring Lake where soil and environmental conditions like ALD, soil temperature, vegetation and SOC were similar and at a dry peat site in Seida. Despite these differences, the averaged contributions of $R_H$ to $R_{eco}$ of 42% at the center and 60% at the rim are in good agreement with those observed at Seida (37 – 64%) and Daring Lake (44 – 64%). Similar contributions were determined from an arctic tussock tundra site where $R_H$ makes up approximately 40% of growing season $R_{eco}$ (Segal & Sullivan, 2014) and from a moist acidic tussock tundra site (Hicks Pries et al., 2013). In contrast to these results, Dorrepaal et al. (2009) determined a substantial higher contribution of $R_H$ to $R_{eco}$ with about 70% in a subarctic peatland.
Table 8: Comparison of daily averaged $R_H$ fluxes. The differences between the sites are most probably caused by substrate availability, with higher $R_H$ fluxes simultaneously to higher SOC contents. All listed fluxes were measured with the closed chamber technique.

<table>
<thead>
<tr>
<th>Location</th>
<th>Tundra type</th>
<th>Period</th>
<th>$R_H$ (g C m$^{-2}$ d$^{-1}$)</th>
<th>Ref</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lena River Delta, RU</td>
<td>polygonal rim/polygonal center</td>
<td>Jul-Sep 2015</td>
<td>0.5 ± 0.1 0.3 ± 0.02</td>
<td>a</td>
</tr>
<tr>
<td>Daring Lake, CA (65°N,</td>
<td>dry heath/mesic birch/wet sedge</td>
<td>Jun-Sep 2004</td>
<td>0.8 ± 0.1 1.8 ± 0.2 0.4 ± 0.1</td>
<td>b</td>
</tr>
<tr>
<td>111°W)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Seida, RU (67°N, 63°E)</td>
<td>dry peat/moist peat/bare peat/shrub tundra heath</td>
<td>Jun-Sep 2008</td>
<td>0.4 ± 0.03 0.6 ± 0.2 1.0 ± 0.04 0.6 ± 0.07</td>
<td>c</td>
</tr>
</tbody>
</table>

a: This study; b: Nobrega and Grogan (2008); c: Biasi et al. (2014)

6.1.4. Spatial variability of heterotrophic respiration fluxes

The partitioning of $R_{eco}$ fluxes revealed a higher contribution of $R_H$ to $R_{eco}$ at the polygonal rim compared to the polygonal center. This difference of the $R_H/R_{eco}$ ratio can be related to differences in vascular plant coverage and moisture conditions between both microsites. Higher GPP as at the center compared to the rim causes also higher rates of $R_A$ and in turn lowers the contribution of $R_H$ to $R_{eco}$. Additionally, anoxic soil conditions due to standing water at the center are not favorable for decomposition of SOM. Furthermore, Nobrega and Grogan (2008) concluded that consistently moderate moisture conditions, as at the rim microsite, promotes fast decomposition of SOM and therefore ensures high $R_H$ rates. In general, $R_H$ fluxes are expected to correlate with SOC contents, where higher SOC contents promote higher $R_H$ fluxes (Biasi et al., 2014, Lavoie et al., 2011). However, SOC contents at the polygonal center were about two-times higher in the upper soil layer compared to the rim (Figure 18 and Figure 19), but $R_H$ fluxes were lower. This relationship is most likely masked by other environmental factors, most dominantly by anoxic conditions at the center and higher abundance of vascular plants. Furthermore, other factors such as soil physical conditions (Schimel et al., 1994) as well as the soil microbial community composition and quantity (Elberling et al., 2008a) can influence the $R_H/R_{eco}$ ratio.

Höfle (2015) reported differences of the SOM composition in the polygonal tundra on Samoylov Island: The SOM of polygonal centers consist of mostly undecomposed
6. Discussion

plant-derived SOM with relatively young $^{14}$C ages, whereas the rims consist of much older, little-decomposed plant material. Based on these findings, a slow microbial SOM degradation at the rim is suggested (Höfle, 2015). However, the differences of the $R_{\text{eco}}$ and $R_{\text{H}}$ fluxes between rim and center from this study lead to an opposing assumption. The fact that at the center, despite higher GPP fluxes, the $R_{\text{eco}}$ and $R_{\text{H}}$ fluxes are lower compared to those from the rim, indicates a higher microbial decomposition rate at the rim. It could be suggested that higher $^{14}$C ages with increasing soil depth, found by Höfle (2015) at the polygonal rim, are because of high decomposition rates of fresh plant material at this microsite. In general, easily decomposable C compounds with young $^{14}$C ages (available from fresh plant material) are mineralized faster as more stable C compounds (von Lützow et al., 2008, Walz et al., 2017), and the amount of more stable C compounds increases with soil depth (Schädel et al., 2014), which could lead to a pronounced age stratification in the soil. In turn, younger $^{14}$C ages at the center might be caused by accumulation of recent plant material due to low decomposition rates in this water-logged environment, observed by low $R_{\text{eco}}$ and $R_{\text{H}}$ fluxes in this study. Furthermore, cryoturbation in the soils of polygonal rims could heave SOM with relatively old $^{14}$C ages upward into soil layers with younger SOM, which would lead to a pronounced age stratification in these soils.

The averaged contribution of $R_{\text{H}}$ to $R_{\text{eco}}$ fluxes during the measurement period in 2015 were 42% at the center and 60% at the rim. Thus, Hypothesis 3, that the contribution of $R_{\text{H}}$ on $R_{\text{eco}}$ is below 50% during the growing season, can be supported for $R_{\text{H}}$ fluxes at the polygonal center, but has to be rejected for $R_{\text{H}}$ fluxes at the rim. Neither at the rim nor at the center was a seasonal trend of the $R_{\text{H}}/R_{\text{eco}}$ ratio observed. This is in contrast to a study from Segal and Sullivan (2014) where the $R_{\text{H}}/R_{\text{eco}}$ ratio increased towards the end of the growing season, most likely due to deepening of the active layer which increases substrate availability for $R_{\text{H}}$ processes. This effect might be missed in this study because of smaller changes in ALD (Figure 15) as well as lower soil temperatures throughout the growing season at the study site compared to other arctic tundra sites. Also other factors like increased substrate availability due to plant senescence and root mortality are suggested to lead to an increased $R_{\text{H}}/R_{\text{eco}}$ ratio towards the end of the growing season in prairie grasslands (Gomez-Casanovas et al., 2012). Whether these factors are of major
The importance in arctic tundra ecosystems remains uncertain as it was recently shown that increased substrate availability via priming has minor effects on \( R_H \) fluxes in the active layer (Walz et al., 2017). Even at the end of the measurement period, considerable GPP fluxes were measured, which in turn means that substantial amounts of plants are still photosynthetically active. Therefore, these active plants are not yet providing dead plant material as substrate to prime microbial decomposition, which would lead to higher \( R_H \) fluxes. Based on this finding, it is concluded that the investigation period was too short to observe significant changes in the \( R_H/R_{eco} \) ratio and an increase of this ratio may be observed later in the year. Therefore, Hypothesis 4, that the \( R_H/R_{eco} \) ratio has a distinct seasonality with higher values towards the end of the season cannot be answered.

Interestingly, significant correlations of the WT fluctuations with the \( R_H/R_{eco} \) ratio and \( R_A \) fluxes (Figure 27), but no correlation between \( R_H \) fluxes and WT were observed. Most likely, the \( R_A \) fluxes are negatively affected by high WT due to submersion of the moss layer and vascular leaf area as submersion can lead to plant stress which reduces productivity and nutrient turnover (Gebauer et al., 1995). \( R_H \) fluxes might be unaffected by variations of high standing water as the decomposition of SOM takes place in deeper parts of the soil, which are permanently water-saturated. Only in times of remarkable low WT the \( R_H/R_{eco} \) ratio increases, which shows low \( R_A \) contribution during these periods. Too low soil moisture contents can limit the growth and productivity of an ecosystem (Chen et al., 2015) as desiccation lowers the photosynthetic activity (Turetsky et al., 2012) and in turn lowers \( R_A \) fluxes. This could lead to the observed increase of the \( R_H/R_{eco} \) ratio simultaneously to a lower water table. These findings show the importance of the WT and VWC for \( R_{eco} \) fluxes. The relationship between \( R_A \) fluxes and WT fluctuations explain the need of WT and VWC for the reproduction of the \( R_{eco} \) fluxes (and not for the \( R_H \) fluxes) over the complete measurement period using the empirical \( Q_{10} \) model.

6.2. CH\(_4\) fluxes

6.2.1. Factors regulating CH\(_4\) fluxes

The CH\(_4\) emissions from the polygonal rim and center in 2015 can partwise be explained with environmental variables (Table 6). In contrast to that, no significant correlation was found between CH\(_4\) emissions from polygonal rim and center in 2014. However, the CH\(_4\)
emissions in 2014 showed a much lower range as in 2015 (Figure 31 and Figure 32) and the measurements are limited to just 17 days in August. This may cause poor correlations with environmental variables as within this rather short period the environmental variables also showed low variations (Figure 12 and Figure 13). The environmental parameter with the highest explanatory power of the CH$_4$ fluxes in 2015 for both microsites was the ALD. Therefore, Hypothesis 5, that the soil temperature has the highest explanatory power of CH$_4$ fluxes, is rejected. However, this finding is in good agreement with other studies who found the ALD to be an important predictor of CH$_4$ emissions (Friborg et al., 2000, van Huissteden et al., 2005). A relationship between ALD and CH$_4$ emissions could indicate a substantial contribution to CH$_4$ emissions from deeper soil layers (Wille et al., 2008). A higher active layer increases the microbial active soil column and are accompanied with higher soil temperatures, which causes higher methanogenic activity (Wagner et al., 2007). In contrast to that, Olefeldt et al. (2013) found no correlation between CH$_4$ emissions and ALD by assembling several studies on CH$_4$ fluxes from arctic ecosystems. However, the relationship of ALD and CH$_4$ emissions is likely to be masked in many ecosystems by decreasing soil temperatures with increasing soil depth, which causes lower microbial activity (Conrad, 1996), even if the microorganisms are adapted to cold conditions (Wagner et al., 2007). In turn, the soil temperature has been identified by many studies to be one of the main environmental parameter controlling CH$_4$ emissions in wetlands (e.g. McEwing et al., 2015, Olefeldt et al., 2013, Wille et al., 2008). In this study, the soil temperature poorly explained CH$_4$ emissions from the polygonal rim, whereas at the center the correlation was slightly better (Table 6). This is in accordance to Olefeldt et al. (2013), who suggested that ecosystems with a WT at or above the soil surface are more sensitive to soil temperature variability than drier ecosystems. It was previously shown that a good correlation between soil temperature and CH$_4$ fluxes only occurs at inundated sites, with a WT distinctly above the soil surface (Nykänen et al., 1998). If the WT is close to the soil surface or even below, as at the polygonal rim and center, the upcoming methane oxidation is suggested to mask the relationship between soil temperature and CH$_4$ emissions (Kutzbach et al., 2004, Zhu et al., 2014). As a result of that, CH$_4$ fluxes can often be related to the WT fluctuations (Marushchak et al., 2016, Olefeldt et al., 2013). However, at polygonal rim and center, WT and VWC had a weak explanatory power of the CH$_4$ emissions in 2015 (Table 6). The effect
of WT fluctuations on CH₄ emissions are described as an on-off switch (Christensen et al., 2003). Some authors suggested that a water table just slightly below the soil surface can cause high CH₄ oxidation rates (Kutzbach et al., 2004, Whalen et al., 1996). Even under water-saturated soil conditions substantial amounts of CH₄ can be oxidized by brown mosses at the study site (Liebner et al., 2011). Furthermore, it was shown that CH₄ emissions are not always well correlated with water table fluctuations and that dry periods can lower CH₄ emissions considerably even after rewetting (Brown et al., 2014). All these findings can lead to a weak correlation between the water table and CH₄ emissions.

Another factor regulating CH₄ production and consumption processes might be given by the pH value. Both methanotrophs and methanogens are known to be neutrophilic (Hanson & Hanson, 1996, Whalen, 2005) and the optimum pH of methane production and oxidation in subarctic peatlands is suggested to be 5.5-7.5 and 5.0-6.5, respectively (Dunfield et al., 1993). On the other hand, Valentine et al. (1994) observed a significant reduction of CH₄ production potential by lowering the pH from 7.0 to 5.5 in a northern fen and Kamal and Varma (2008) suggested more acidic pH values as favorable for the growth of methanotrophs. The comparison between the pH values of the porewater analysis from the center (Figure 20) and CH₄ fluxes (Figure 32) reveals a linear relationship between emissions and pH. Here, highest emissions were measured during times when the pH was most acidic ($R^2 = 0.73$ – data not shown). This might give evidence for an impact of soil porewater pH on CH₄ emissions. However, as data on porewater pH are very scarce during the measurement period, the assumed relationship cannot be fully investigated.

6.2.2. CH₄ fluxes in comparison with other arctic tundra sites

The averaged CH₄ fluxes were compared with reported CH₄ fluxes from arctic tundra sites that are similar in vegetation and soil composition (Table 9). In general, the comparison revealed large differences of the CH₄ emissions between study sites and sampling years. These differences might be caused due to different environmental conditions, different sampling strategies and measurement periods. For instance, methodological differences are the chamber size and equipment as well as closure times, sampling periods and
frequencies. Nevertheless, as all listed studies were conducted in the polygonal tundra, or at least in ecosystems that are similar to the polygonal tundra, this comparison provides a profound overview about CH₄ fluxes from these ecosystems.

The determined CH₄ fluxes from this study are at the lower end in comparison with CH₄ fluxes from other studies (Table 9). A set of studies from wet tundra sites reported more than four-fold higher CH₄ fluxes compared to the polygonal center CH₄ emissions from this study in 2015 (Bartlett et al., 1992, McEwing et al., 2015, Ström et al., 2012, van Huissteden et al., 2005, Vaughn et al., 2016). Studies of CH₄ emissions from relatively dry tundra sites also determined partially distinctly higher CH₄ fluxes compared to the polygonal rim of this study (van Huissteden et al., 2005, Vaughn et al., 2016). Only a few studies determined CH₄ emissions from wet tundra (Heikkinen et al., 2002, Marushchak et al., 2016) and dry tundra ecosystems (Davidson et al., 2016) that were lower as CH₄ emissions from this study.

The rather low CH₄ emissions from the polygonal tundra on Samoylov Island can be related to low SOM contents and permafrost temperatures. Furthermore, the soils of the polygonal tundra on Samoylov Island are characterized by a sandy texture, which is not a favorable habitat for methane producing microbes (Wagner et al., 1999). The low permafrost temperature at the study site (Boike et al., 2013, Romanovsky et al., 2010) is suggested to lead to low CH₄ production rates as methanogenesis is controlled by temperature (Whalen, 2005). The soil organic matter contents are of major importance for CH₄ emissions. The comparison between CH₄ fluxes from this and other studies (Table 9) shows that highest CH₄ emissions were measured in organic soils (Bartlett et al., 1992, McEwing et al., 2015, Ström et al., 2012, Vaughn et al., 2016), while CH₄ emissions from mineral soils are generally lower. This is in good agreement with Knoblauch et al. (2015) who suggested that low SOM contents in the polygonal tundra of the Lena River Delta are a reason for comparatively low CH₄ emissions. Furthermore, SOM contents are found to be positively correlated with CH₄ emissions across several tundra ecosystems (Christensen et al., 1995).
Table 9: Comparison of daily averaged \( \text{CH}_4 \) fluxes measured at various arctic tundra sites. The sites are similar in vegetation and land cover class. All listed fluxes were measured with the closed chamber technique. The soils are categorized as organic if SOC contents are > 20%. *: standard error not available

<table>
<thead>
<tr>
<th>Location</th>
<th>Tundra type</th>
<th>Period</th>
<th>Soil</th>
<th>( \text{CH}_4 ) flux (ng m(^{-2}) d(^{-1}))</th>
<th>Ref</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lena River Delta, RU (72°N, 127°E)</td>
<td>polygonal center polygonal rim polygonal center polygonal rim</td>
<td>August 2014</td>
<td>mineral mineral mineral</td>
<td>38.4 ± 2.2 1.7 ± 0.2 24.1 ± 2.9 1.7 ± 0.6</td>
<td>a</td>
</tr>
<tr>
<td>Yukon Delta, US (60°N, 162°W)</td>
<td>wet tundra dry upland tundra</td>
<td>Jul-Aug 1988</td>
<td>organic</td>
<td>143.6 ± 31.1 2.3 ± 1.1</td>
<td>b</td>
</tr>
<tr>
<td>Lek Vorkuta, RU (67°N, 63°E)</td>
<td>wet flark</td>
<td>Jun-Sep 1999</td>
<td>organic</td>
<td>6.0 ± *</td>
<td>c</td>
</tr>
<tr>
<td>Lena River Delta, RU (72°N, 127°E)</td>
<td>polygonal center polygonal rim</td>
<td>Aug 1999</td>
<td>mineral mineral</td>
<td>28.0 ± 5.4 4.3 ± 0.8</td>
<td>d</td>
</tr>
<tr>
<td>Tiksi, RU (72°N, 130°E)</td>
<td>wet sedge tundra</td>
<td>Jul-Aug 1993</td>
<td>organic</td>
<td>46.3 ± *</td>
<td>e</td>
</tr>
<tr>
<td>Lena River Delta, RU (72°N, 127°E)</td>
<td>polygonal centers polygonal rim</td>
<td>Jul-Sep 2006</td>
<td>mineral mineral</td>
<td>86.2 ± 25 4.9 ± 10</td>
<td>f</td>
</tr>
<tr>
<td>Zackenberg, GL (74°N, 20°W)</td>
<td>sedge-dom. wet tundra</td>
<td>Jun-Aug</td>
<td>organic</td>
<td>130.6 ± 13</td>
<td>g</td>
</tr>
<tr>
<td>Chokurdakh, RU (71°N, 147°E)</td>
<td>wet (pol.) tundra dry (pol.) tundra</td>
<td>Jul 2004</td>
<td>organic</td>
<td>171.4 ± * 4.3 ± *</td>
<td>h</td>
</tr>
<tr>
<td>Lena River Delta, RU (72°N, 127°E)</td>
<td>polygonal center polygonal rim</td>
<td>Jun-Sep 1999</td>
<td>mineral mineral</td>
<td>53.2 ± 8.7 4.7 ± 2.5</td>
<td>i</td>
</tr>
<tr>
<td>Seida, RU (67°N, 63°E)</td>
<td>sedge-dom. tundra</td>
<td>Jul-Oct 2007 May-Oct 2008</td>
<td>organic</td>
<td>10.1 ± 1.0 6.8 ± 1.3</td>
<td>j</td>
</tr>
<tr>
<td>Barrow, US (71°N, 157°W)</td>
<td>wet sedge tundra</td>
<td>Jul-Sep 2013</td>
<td>organic</td>
<td>108.5 ± 10.8</td>
<td>k</td>
</tr>
<tr>
<td>Barrow, US (71°N, 157°W)</td>
<td>wet sedge tundra dry tundra</td>
<td>Jun-Aug 2014</td>
<td>organic</td>
<td>40.3 ± 48.5 1.4 ± 3.6</td>
<td>l</td>
</tr>
<tr>
<td>Barrow, US (71°N, 157°W)</td>
<td>polygonal center polygonal rim</td>
<td>Jul-Oct 2013</td>
<td>organic</td>
<td>122.2 ± 63.6 24.2 ± 18.0</td>
<td>m</td>
</tr>
</tbody>
</table>

a: This study; b: Bartlett et al. (1992); c: Heikkinen et al. (2002); d: Kutzbach et al. (2004); e: Nakano et al. (2000); f: Sachs et al. (2010); g: Ström et al. (2012); h: van Huissteden et al. (2005); i: Wagner et al. (2003); j: Marushchak et al. (2016); k: McEwing et al. (2015); l: Davidson et al. (2016); m: Vaughn et al. (2016)

6.2.3. \( \text{CH}_4 \) fluxes in comparison with earlier studies from Samoylov Island

Beside this study, three other studies focused on \( \text{CH}_4 \) emissions from the polygonal tundra on Samoylov Island on the microscale (Kutzbach et al., 2004, Sachs et al., 2010, Wagner et al., 2003). Between these studies very large differences of \( \text{CH}_4 \) emissions from polygonal center and rim were determined: Kutzbach et al. (2004) found \( \text{CH}_4 \) fluxes from a polygonal center that were within the same range as determined in this study in 2015, even though the \( \text{CH}_4 \) emissions were measured just in August. On the other hand, Wagner et al. (2003) and Sachs et al. (2010) found the \( \text{CH}_4 \) fluxes from a polygonal center to be
6. Discussion

substantial higher compared to CH\textsubscript{4} emissions from this study (Table 9). Additionally, the CH\textsubscript{4} emissions from polygonal rims were within the same range between the three studies, but differ significantly to determined fluxes from this study. These differences most likely display the large temporal and spatial variability of CH\textsubscript{4} fluxes on microsite even within the same study site. Variable biotic and abiotic conditions between the study years most likely cause these differences. For instance, different pH values are suggested to could have an impact on CH\textsubscript{4} production and oxidation rates (see section 6.3.1.). While the soil porewater at the center from this study was slightly acidic (Figure 20), Wagner et al. (2003) measured pH values between 7.4 and 7.9. These slightly alkaline pH values could increase the rates of methanogenesis and decrease the methane oxidation (Kamal & Varma, 2008, Valentine et al., 1994), which would partly explain higher measured CH\textsubscript{4} emissions from Wagner et al. (2003) compared to those from this study. Furthermore, disparities in the sampling procedure as well as the calculation of the CH\textsubscript{4} fluxes might explain the fluxes differences between years. The use of different regression models for the flux calculation can lead to substantial differences of the fluxes (Kutzbach et al., 2007a). While in this study as well as in the study from Kutzbach et al. (2004), the CH\textsubscript{4} emissions were calculated using a linear regression model, Sachs et al. (2010) used a non-linear regression model for the flux calculation (Figure 33). It is well known that there is a non-linear nature of gas concentration evolution over time in closed chambers due to disturbance of the gas concentration gradient during the chamber measurement (Hutchinson et al., 2000, Livingston et al., 2006). Nevertheless, as mentioned in the section 4.5.1., a linear regression model can produce a better flux estimate for a non-linear concentration-over-time curve than a non-linear regression model (Görres et al., 2014).

Furthermore, initial pressure shocks could lead to an overestimation of CH\textsubscript{4} emissions, but it remains uncertain, if they can explain the discrepancies of CH\textsubscript{4} emissions from this study to those from Wagner et al. (2003) and Sachs et al. (2010). At least initial pressure shocks could possibly explain the distinct non-linear flux evaluation of CH\textsubscript{4} fluxes measured by Sachs et al. (2010) (Figure 33). In general, if no attention is paid to initial pressure shocks, the CH\textsubscript{4} fluxes can be substantially overestimated even if the flux calculation was done with a linear regression model (Christiansen et al., 2011). This especially holds true for flux
estimates based on measurements, where samples are taken in intervals of up to several minutes or based on one-point measurements (like Wagner et al. (2003)), as it is possible that initial pressure shocks are missed.

**Figure 33:** Examples of different concentration-over-time curves of CH$_4$ fluxes. Panel (a) shows the concentration-over-time curve from CH$_4$ fluxes measured in this study. Panel (b) is modified after figure 4 in Sachs et al. (2010). The distinct non-linearity of the CH$_4$ fluxes measured by Sachs et al. (2010) might be an artefact of the chamber placement, called initial pressure shocks. Note the different scale of both axis between panel (a) and (b).

The measured CH$_4$ fluxes in 2014 and 2015 from this study showed no difference between years in CH$_4$ emissions at the polygonal rim (Table 9). In contrast to that, the CH$_4$ emissions at the center were distinctly higher in 2014 compared to 2015. This difference might demonstrate the inter-annual variability of CH$_4$ emissions at the polygonal tundra. A possible reason for higher CH$_4$ emissions in 2014 is the higher air temperature in August 2014 (about 2 °C higher as in August 2015). As mentioned in section 6.3.1., the temperature is assumed to be a major driver of CH$_4$ fluxes and could therefore be a possible explanation for CH$_4$ flux differences between the studies in 2014 and 2015. If the measured CH$_4$ emissions in 2014 are compared to a period of same length in 2015, where highest CH$_4$ emissions occurred, the gap between the emissions decreases: Between 24 August and 13 September 2015 a daily averaged CH$_4$ emission of $0.37 \pm 0.08$ µg m$^{-2}$ s$^{-1}$ was measured, which is close to the mean emissions of $0.45 \pm 0.03$ µg m$^{-2}$ s$^{-1}$ in 2014. It might be possible that the measurement period in 2014 only contains the peak of CH$_4$ emissions, which occurred later in 2015. This would result in higher mean CH$_4$ emissions in comparison to the same period in 2015, where the peak emissions were measured about three weeks later (Figure 32). However, the measurement period in 2014 was too
short to gain a sufficient insight into the seasonality of the CH emissions for reliable estimates of the CH$_4$ flux dynamics in 2014.

### 6.2.4. CH$_4$ flux differences on the microscale

A considerable difference between CH$_4$ emissions of the polygonal center and rim was observed in this study (Figure 31 and Figure 32). These differences can be attributed to a set of differing environmental and physical controls influencing CH$_4$ production, oxidation as well as release mechanisms. One of the main factors leading to a substantial difference between microsites is the water table (e.g., Harazono *et al*., 2006, Marushchak *et al*., 2016, Olefeldt *et al*., 2013). With a water table close to the soil surface, the methane production zone at the center has a higher thickness as at the rim, where the water table is just a few centimeters above the permafrost table (Figure 16). This difference causes most likely higher CH$_4$ production rates at the polygonal center compared to the rim. Due to non-water-saturated conditions at the rim, most parts of the active layer are well aerated, which means a higher thickness of the methane oxidation zone in relation to the center. Additionally, due to non-water-saturated conditions, ebullition as transport mechanism of CH$_4$ molecules from the soil to the atmosphere is lacking at the rim.

The vegetation composition is another factor that possibly explains the CH$_4$ flux differences between the microsites. At the polygonal center, higher growth forms and abundance of *Carex Aquatilis* were observed compared to the rim (Table 4). It is well established that the abundance of sedges is an important factor controlling CH$_4$ emissions (King *et al*., 1998, Verville *et al*., 1998) as they provide a transport mechanism for CH$_4$ where oxic zones of the soil are bypassed, namely plant-mediated transport (Kutzbach *et al*., 2004). Also the height of vascular plants is suggested to cause higher CH$_4$ emission rates due greater CH$_4$ transport capacities as well as enhancement of substrate supply for methanogenesis with taller plants (von Fischer *et al*., 2010). Substrate availability is identified as one of the main drivers of CH$_4$ fluxes (Christensen *et al*., 2003, McEwing *et al*., 2015). In deeper soil layers at the polygonal center, the substrate availability, higher SOC and N values serve here as proxy for higher substrate availability, was distinctly higher as at the rim (Figure 18 and Figure 19). On the other hand, vascular plants are transporting oxygen to their roots which supports methanotrophic activity in the surrounding soil.
(Conrad, 1996, Harazono et al., 2006), but this effect is suggested to play a minor role at the center due to presumably high CH$_4$ production rates in the soil. There is evidence that in mosses of water-logged soils higher methanotrophic activity takes place compared to unsaturated soils (Vecherskaya et al., 1993), probably due to symbiotic-like relations of mosses and methanotrophs as in brown mosses (Liebner et al., 2011). These findings lead to the suggestion that methanotrophic activity is higher at the polygonal center as on the rim, lowering the difference of CH$_4$ emissions between the microsites. However, this effect might has just a low impact on total CH$_4$ fluxes as in water-saturated soils the CH$_4$ production rate exceed the rate of methane oxidation by far. Furthermore, it is suggested that the higher ALD at the center causes higher CH$_4$ emissions. An enlargement of the ALD is accompanied by an increase of the methane production zone, which is supported by the correlation of ALD with measured CH$_4$ emissions from rim and center (Table 6). In contrast to that, Olefeldt et al. (2013) found no correlation between CH$_4$ emissions and active layer depth by compiling various studies on CH$_4$ fluxes across permafrost landscapes. The different dependencies of CH$_4$ fluxes on environmental parameters across different arctic tundra sites highlight the complicated the determination of their individual impact on CH$_4$ production, oxidation and transport mechanisms.

The averaged CH$_4$ emissions at the polygonal rim show that this microsite acts as net source for atmospheric CH$_4$ (Table 9). Furthermore, not a single measured CH$_4$ flux showed a net uptake of CH$_4$ neither at the rim, nor at the center (Figure 31 and Figure 32). Therefore, Hypothesis 6, that the polygonal rim act as sink for atmospheric CH$_4$ is rejected. This is in contrast to other studies who found substantial net CH$_4$ uptake rates in non-water-saturated high arctic tundra soils in Zackenberg, Greenland (Jørgensen et al., 2015) and antarctic tundra soils (Zhu et al., 2014). These CH$_4$ uptake rates are positively correlated to soil temperatures. In this study, however, no significant correlation between CH$_4$ emissions and soil temperature was observed, but a correlation might be masked due to simultaneously increasing methanogenesis with higher soil temperatures. It is suggested that substrate availability and soil temperatures are of major importance for the differences of CH$_4$ sink strengths between these ecosystems. The SOC contents of dry tundra soils in Greenland are lower than those at the polygonal rim (Figure 19 and Elberling et al. (2008b)). This most likely lowers the methanogenesis in these soils as a low
substrate availability is not favorable for CH₄ production (Ström et al., 2012). The soil temperatures at the polygonal rim were higher than soil temperatures at antarctic tundra soils (Figure 15 and Zhu et al. (2014)). As the CH₄ production has a lower temperature dependence than the CH₄ oxidation (Dunfield et al., 1993), higher soil temperatures in the oxic soil layers lead to a higher increase of CH₄ oxidation rates compared to the increase of CH₄ production. As a result, the lower substrate availability in Greenland soils and higher soil temperatures in antarctic soils lead to a higher impact of CH₄ oxidation on the CH₄ fluxes, which can cause in particular cases a net CH₄ sink. In contrast to this, the rather high substrate supply and low soil temperature at the polygonal rim are not promoting the CH₄ sink function of this microsite.

6.3. CO₂ and CH₄ budgets in the polygonal tundra

The modeled CO₂ and mean CH₄ fluxes from polygonal rim and center were used to calculate the CO₂ and CH₄ budget of the measurement period in 2015 of the EC footprint area of the EC system and the polygonal tundra of Samoylov Island (Table 10). Based on the surface classification from Muster et al. (2012), the polygonal rim makes up 65% of the polygonal tundra of Samoylov Island, whereas the center represent 19%. Open water bodies make up 16% of the polygonal tundra. As CO₂ and CH₄ fluxes from this microsite are missing in this study, the estimates of CO₂ were taken from Abnizova et al. (2012) and estimates of CH₄ from Knoblauch et al. (2015).

Table 10: Calculation of the CO₂ and CH₄ budget for the measurement period in 2015. The budgets were calculated for the polygonal tundra on Samoylov Island.

<table>
<thead>
<tr>
<th>Land cover class</th>
<th>Coverage (%)</th>
<th>CO₂ flux (g CO₂ m⁻² d⁻¹)</th>
<th>Total CO₂ flux (Kg CO₂ ha⁻¹ d⁻¹)</th>
<th>CH₄ flux (mg CH₄ m⁻³ d⁻¹)</th>
<th>Total CH₄ flux (Kg CO₂-equ ha⁻¹ d⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dry tundra</td>
<td>65</td>
<td>-2.3</td>
<td>-14.8</td>
<td>1.7</td>
<td>0.4</td>
</tr>
<tr>
<td>Wet tundra</td>
<td>19</td>
<td>-5.9</td>
<td>-11.2</td>
<td>25.5</td>
<td>1.7</td>
</tr>
<tr>
<td>Water</td>
<td>16</td>
<td>6.8b</td>
<td>10.9</td>
<td>26.0c</td>
<td>1.4</td>
</tr>
<tr>
<td>Total</td>
<td>100</td>
<td>-</td>
<td>-15.3</td>
<td>-</td>
<td>3.4</td>
</tr>
<tr>
<td>EC fluxes</td>
<td>-</td>
<td>-</td>
<td>-24.6b, -21.0d, -15.1e</td>
<td>-</td>
<td>6.2b, 6.0f, 6.5g</td>
</tr>
</tbody>
</table>

a: Muster et al. (2012); b: Abnizova et al. (2012); c: Knoblauch et al. (2015); d: Kutzbach et al. (2007b); e: Runkle et al. (2013); f: Wille et al. (2008); g: Sachs et al. (2008); h: Kutzbach et al. unpublished.

The calculated total CO₂ budget from the individual microsites polygonal rim, center and ponds is in agreement with estimates of CO₂ fluxes calculated with EC measurements.
during the same period (Kutzbach et al., unpublished). With -15.3 Kg CO$_2$ ha$^{-1}$ d$^{-1}$, the calculated CO$_2$ budget is also in agreement to EC footprint estimates from other years (Kutzbach et al., 2007b, Runkle et al., 2013). The differences of EC footprint estimates between years demonstrate the inter-annual variability of the CO$_2$ fluxes at the study site, which are caused, among others, by differing meteorological conditions. For instance, variety of PAR can lead to different GPP fluxes between years, as PAR is the most important driver of photosynthesis. During the growing season in 2003, the average air temperature in July was about 2 °C higher than in 2015 (Figure 16) and total rainfall was twice as high as during the measurement period in 2015 (Figure 15) (Kutzbach et al., 2007b). These differences can cause different CO$_2$ NEE fluxes as temperature and precipitation rates affect all of the NEE underlying processes (e.g. Hobbie et al., 2002, Luo & Zhou, 2006, Schlesinger, 2013). Furthermore, the estimates of the outgassing CO$_2$ from polygonal ponds are related to high variabilities (Abnizova et al., 2012), which could cause a strong bias of the estimate.

On polygonal tundra-scale, the polygonal rim displays a stronger CO$_2$ sink as the center, which supports Hypothesis 7. This is remarkable as almost three-fold lower net CO$_2$ uptake fluxes were measured at the rim compared to the center (Table 5). Therefore, small changes of the CO$_2$ NEE at the rims can have large impacts on the CO$_2$ budget of the polygonal tundra. This highlights the importance of polygonal rims for the CO$_2$ sink strength.

Considerably lower CH$_4$ emissions of the polygonal tundra area were calculated with chamber flux estimates from this study compared to EC measurements from other investigation periods (Sachs et al., 2008, Wille et al., 2008). Also a first estimate of CH$_4$ emissions from EC measurements (Kutzbach et al., unpublished) revealed higher CH$_4$ emissions than the summed CH$_4$ fluxes for the polygonal tundra from the single microsites rim, center and ponds. This difference most likely reflects the distinct spatial variability of CH$_4$ fluxes in the polygonal tundra and highlights the difficulties of reliable estimates of CH$_4$ emissions from these landscapes. The WT, one of the main drivers of CH$_4$ production, is highly variable on the spatial scale; For instance, at the polygonal center from the study site the water table is partly below the soil surface, which causes high rates of CH$_4$ oxidation in the upper soil layer. Most likely these conditions are highly heterogeneous
across polygonal centers on Samoylov Island. This could lead to a substantial bias in the budget calculation. Furthermore, the CH$_4$ emission value from the ponds used for the calculation is representing a mean of measurements from an open water body and a pond margin (Knoblauch et al., 2015). This value is related to uncertainties as it remains unclear how pond margins and open water bodies are distributed in the polygonal tundra. Therefore, it might be possible that a substantial source is missed or underestimated in the CH$_4$ budget calculation based on the fluxes from the single microsites.

Considering the 34-fold higher GWP of CH$_4$ on a 100-year timescale compared to CO$_2$ (Myhre et al., 2013), the CH$_4$ emissions might display an important contributor to the overall greenhouse gas balance of the polygonal tundra. The conversion of CH$_4$ emissions into CO$_2$-equivalents revealed that the net uptake of atmospheric CO$_2$ of -15.3 kg CO$_2$ ha$^{-1}$ d$^{-1}$ was about five-fold higher as the CH$_4$ emissions of 3.4 Kg CO$_2$-eq ha$^{-1}$ d$^{-1}$. Therefore, the polygonal tundra of Samoylov Island displayed a robust sink for atmospheric carbon in 2015. If the same calculation is conducted with CO$_2$ and CH$_4$ fluxes from 2014, the polygonal tundra represents a net source for atmospheric carbon as CO$_2$ fluxes are with 3.7 Kg CO$_2$ ha$^{-1}$ d$^{-1}$ a small and the CH$_4$ fluxes with 8.1 Kg CO$_2$-eq ha$^{-1}$ d$^{-1}$ a large source. However, CH$_4$ and CO$_2$ flux measurements in 2014 are restricted to just one month and it might be possible that those fluxes show a different trend over the complete growing season. Nevertheless, this finding emphasizes the inter-annual variability of C dynamics between soils and atmosphere in the polygonal tundra.
7. Conclusion and Outlook

This study examined the contributions of GPP, $R_{eco}$, $R_H$ and $R_A$ to CO$_2$ NEE fluxes on the microscale in the polygonal tundra. Both investigated microsites represent a sink for atmospheric CO$_2$ during the measurement periods July to September 2015 and July to August 2014. The sink function is more pronounced at the polygonal center compared to the polygonal rim site (Hypothesis 1 is confirmed) and was stronger in 2015 than in 2014. It is concluded, that the difference in CO$_2$ fluxes between the two microsites occurs mainly due to $R_{eco}$ being suppressed under the water-saturated conditions and not primarily because of higher GPP at the center (supports Hypothesis 2). This assumption holds also true for $R_H$ fluxes, which represent the first in situ measurements of $R_H$ fluxes over almost a complete growing season from a Russian arctic tundra site. The substantial differences identified in NEE between the two investigated microsites show the importance of microscale measurements for reliable estimates of CO$_2$ surface-atmosphere fluxes from arctic tundra sites and highlight the important role of soil moisture conditions on CO$_2$ fluxes.

The contribution of $R_H$ to $R_{eco}$ fluxes differs between the microsites. At the polygonal center the average contribution of $R_H$ to $R_{eco}$ is 42% (Hypothesis 3 is supported), while at the polygonal rim the average contribution of $R_H$ is comparatively higher at 60% (Hypothesis 3 is rejected). This difference is most likely related to differences in vascular plant coverage and soil moisture conditions. The $R_H/R_{eco}$ ratio was not found to increase towards the end of the growing season, and it is concluded that the measurement period was too short to observe any seasonality in the $R_H/R_{eco}$ ratio (Hypothesis 4 cannot be examined). Instead, the $R_H/R_{eco}$ ratio and the $R_A$ fluxes were found to correlate with the WT at the polygonal center. This might be an effect of low $R_A$ fluxes during times of high WT due to the submersion of vascular plants, while during times of low WT, the $R_H$ gains more importance on the $R_H/R_{eco}$ ratio due to well-aerated soils and mosses could desiccate, resulting in low $R_A$ fluxes.

In the future Arctic, rainfall is predicted to increase (Christensen et al., 2013) and this work shows for the polygonal tundra on Samolyov Island that high levels of soil moisture conditions cause this environment to function as a stronger CO$_2$ sink. Hereby, $R_A$ and $R_H$
fluxes respond differently to changing moisture conditions. On the other hand, the modeling of CO\(_2\) fluxes revealed that both R\(_H\) and R\(_A\) fluxes correlate positively with rising temperatures, although Hicks Pries et al. (2015) have shown a different response of these fluxes to warming in other arctic tundra ecosystems. Therefore, it remains uncertain whether future climate change will cause the polygonal tundra to act as stronger CO\(_2\) sink or if it will turn into a CO\(_2\) source. This work shows that the hydrological conditions are of major importance for R\(_A\) and R\(_H\) fluxes. Therefore, it is recommended that future studies determining partitioned CO\(_2\) fluxes from arctic tundra ecosystems should focus on the role of hydrological conditions as driver of these fluxes to obtain an in-depth insight into this relationship.

In order to determine the individual impacts of hydrological conditions and temperature on the R\(_H\) and R\(_A\) fluxes, it would be useful to perform both warming and wetting experiments under field conditions. So far, a number of studies have determined the temperature response of NEE, GPP, and R\(_{eco}\) fluxes in arctic ecosystems with warming experiments (e.g. Frey et al., 2008, Natali et al., 2011, Voigt et al., 2016), however, much less research has focused on the response of R\(_A\) and R\(_H\) fluxes to increased temperatures (Hicks Pries et al., 2015). Wetting experiments in arctic tundra ecosystems to determine the individual response of R\(_A\) and R\(_H\) fluxes to changing hydrological conditions are lacking so far, despite their importance as highlighted in this study.

This study also determined CH\(_4\) fluxes from different microsites in the polygonal tundra. The measured CH\(_4\) emissions are rather low in comparison to arctic tundra sites with organic-rich soils, which highlights the importance of substrate availability for CH\(_4\) production. The rather low CH\(_4\) emissions found in this study in comparison to other studies from the same site (Samoylov Island) highlight not only the high temporal and spatial variability of CH\(_4\) emissions, but also the importance of an accurate measurement procedure. ALD was the environmental parameter with the highest explanatory power of CH\(_4\) fluxes from the polygonal center and rim (Hypothesis 5 is rejected). In contrast to other studies, the dry polygonal rim was not found to function as a CH\(_4\) sink (Hypothesis 6 rejected). It is concluded that low soil temperatures (causing low CH\(_4\) oxidation rates) and high substrate availability (causing high CH\(_4\) production rates) were the main factors causing the polygonal rim to function as a source of CH\(_4\). These findings show the varying
impact of different environmental conditions in arctic tundra ecosystems on CH$_4$ fluxes, which complicates estimates of total CH$_4$ emissions from arctic tundra landscapes. Future studies conducting CH$_4$ flux measurements on the microsite scale across a range of arctic tundra landscapes are required to better estimate the recent contribution of CH$_4$ fluxes to the global C cycle and the future response of these fluxes to global warming.

The calculation of CO$_2$ and CH$_4$ budgets for the polygonal tundra on Samoylov Island revealed that the polygonal rims act as a stronger net CO$_2$ sink than the polygonal centers (Hypothesis 7 is supported). Considering the GWP of CH$_4$ on a 100-year timescale, the CH$_4$ emissions from the polygonal tundra are five-fold lower than its net CO$_2$ uptake. However, on the shorter term, calculating the budget with the fluxes measured in 2014 revealed that this ecosystem acted as net source for atmospheric C over a period of one month, which highlights the pronounced inter-annual variability of these fluxes. Furthermore, the observed differences in CH$_4$ emissions and partitioned CO$_2$ fluxes between the microsites clearly show that modelling approaches of C budgets on larger scales should always be supported by microscale measurements to take the pronounced spatial heterogeneity of arctic tundra ecosystems and its impact on C fluxes into account.

In order to gain a quantitative understanding of the contribution of greenhouse gas emissions from arctic ecosystems to global warming, future studies should also focus on other greenhouse gases in addition to CO$_2$ and CH$_4$. For instance, soil warming increases N$_2$O emissions from arctic tundra sites (Voigt et al., 2016). Furthermore, little effort has been made so far in determining methyl halide fluxes in arctic tundra ecosystems. This is critical as it was shown that several arctic and subarctic ecosystems emit considerable amounts of chloroform (CHCl$_3$) (Johnsen et al., 2016) as well as methyl chloride (CH$_3$Cl) (Hardacre et al., 2009). Both N$_2$O and methyl halides emissions can be of major importance due to their much higher GWP compared to CO$_2$. 

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(Tim Eckhardt)