Lateral expansion and carbon exchange of a boreal peatland in Finland resulting in 7000 years of positive radiative forcing

Paul J. H. Mathijssen1, Noora Kähkölä2, Juha-Pekka Tuovinen3, Annalea Lohila3, Kari Minkkinen4, Tuomas Laurila4, and Minna Väliranta1

1Department of Environmental Sciences, University of Helsinki, Helsinki, Finland, 2Department of Geosciences and Geography, University of Helsinki, Helsinki, Finland, 3Atmospheric Composition Research, Finnish Meteorological Institute, Helsinki, Finland, 4Department of Forest Sciences, University of Helsinki, Helsinki, Finland

Abstract Data on past peatland growth patterns, vegetation development, and carbon (C) dynamics during the various Holocene climate phases may help us to understand possible future climate-peatland feedback mechanisms. In this study, we analyzed and radiocarbon dated several peat cores from Kalevansuo, a drained bog in southern Finland. We investigated peatland succession and C dynamics throughout the Holocene. These data were used to reconstruct the long-term atmospheric radiative forcing, i.e., climate impact of the peatland since initiation. Kalevansuo peat records revealed a general development from fen to bog, typical for the southern boreal zone, but the timing of ombrotrophication varied in different parts of the peatland. Peat accumulation patterns and lateral expansion through paludification were influenced by fires and climate conditions. Long-term C accumulation rates were overall lower than the average values found from literature. We suggest the low accumulation rates are due to repeated burning of the peat surface. Drainage for forestry resulted in a nearly complete replacement of typical bog mosses by forest species within 40 years after drainage. The radiative forcing reconstruction suggested positive values (warming) for the first ~7000 years following initiation. The change from positive to negative forcing was triggered by an expansion of bog vegetation cover and later by drainage. The strong relationship between peatland area and peat type with radiative forcing suggests a possible feedback for future changing climate, as high-latitude peatlands may experience prominent regime shifts, such as fen to bog transitions.

1. Introduction

Northern peatlands are efficient long-term carbon (C) sequestering ecosystems. Globally, they have accumulated approximately 500 (±100) Pg C (Pg = 10^15 g) during the Holocene (the last ~11,700 years) [Loisel et al., 2014], which is equivalent to ~30% of global soil C, and nearly equal to the preindustrial atmospheric C reservoir [Yu, 2012]. The uptake of atmospheric carbon dioxide (CO_2) creates a negative radiative forcing; i.e., long-term peatland development has a cooling effect on the atmosphere. However, the net climate effect depends on the level of simultaneous methane (CH_4) emissions that induce a positive radiative forcing (warming) [Frolking and Roulet, 2007; Korhola et al., 2010; Yu, 2011]. CH_4 emissions represent up to 25% of the net ecosystem C balance of peatlands [Limpens et al., 2008]. During their development, peatlands grow in vertical and horizontal directions [Korhola, 1994], and accordingly, the C fluxes and climate forcing vary through space and time in relation to the successional stage of the peatland [Drewer et al., 2010; Leppälä et al., 2011; Tolonen and Turunen, 1996] and to changes in climatic factors [Dorrepaal et al., 2009; Fan et al., 2013; Gorham, 1991]. Climate conditions are predicted to change considerably in the future toward higher temperatures and more frequent extreme weather, although how these changes will impact on peatland functioning is not clear [Gong et al., 2013; McGuire et al., 2009]. Past data on peatland-climate interactions may help us to understand future peatland C dynamics and the related climate forcing. Of particular interest is the mid-Holocene time period between approximately 8000 and 5000 years before present (B.P.). In Fennoscandia, this period was approximately 2°C warmer and drier than the present climate [Seppä et al., 2005; Väliranta et al., 2005] and can thus be used as a potential future climate analogue.

In general, long-term vertical peat accumulation results in a successional change where hydrological conditions and vegetation structure change over the course of time [Hughes, 2000; Tuittila et al., 2013]. In Finland, early stage peatlands are predominantly characterized by sedge-dominated fen vegetation, high
CH4 emissions [Leppälä et al., 2011], and relatively slow C accumulation rates [Tolonen and Turunen, 1996]. In the southern boreal zone, peatlands that were initially fens have undergone a fen-to-bog-transition, i.e., ombrotrophication, in contrast to more northerly regions where this transition has not occurred [Vääränta et al., 2016]. During ombrotrophication the peat surface becomes disconnected from minerogenic water sources [Rydin and Jeglum, 2013]. In boreal mires peat growth commonly accelerates after ombrotrophication, and this is related to an increased distance between peat surface and water table, an increase in the proportion of Sphagnum and dwarf shrubs and a decrease in pH and decomposability of plant litter; these factors subsequently result in more rapid C sequestration [Tolonen and Turunen, 1996]. Ombrotrophication may occur asynchronously across a single peatland [e.g., Glaser et al., 1981]. Furthermore, lateral peatland expansion not only increases the surface area of C uptake but may also increase CH4 emissions [Korhola et al., 1996].

The peatland at Kalevansuo in southern Finland—examined in this study—is not in a natural state, as it has been drained for forestry. In general, drainage for forestry results in a lowering of the water table through ditching, and this causes rapid changes to vegetation, namely, an increased coverage of trees, forest mosses, and shrubs at the expense of sedges, mire shrubs, and Sphagnum species [Laine et al., 1995]. The lower water table generally results in subsidence of the peat surface and increased peat bulk density [Minkkinen and Laine, 1998]. Drainage affects the C balance by changing, for example, plant productivity, litter quality, decomposition rates, and the residence time of organic matter in aerated conditions [Laiho, 2006]. Typically, these changes result in a decrease in CH4 emissions [Nykänen et al., 1998]. However, whether drained peatlands function as C sinks or sources in the long term is dependent on peatland type, climate, and the extent of change in the water level [Ojanen et al., 2013].

Approximately 4% of the northern peatland area is estimated to have been drained for forestry [Gorham, 1991; Paavilainen and Pääväänen, 1995]. This type of peatland management is especially prevalent in Finland, where ~57,000 km² of peatlands were drained for forestry during the last century, equivalent to 55% of the total original peatland area [Turunen, 2008]. Here drainage has often increased the C uptake [Minkkinen and Laine, 1998; Minkkinen et al., 2002], and correspondingly, it has been estimated that the widespread drainage activities in Finland have had a cooling effect on the atmosphere [Minkkinen et al., 2002], although this effect is partly negated by changes in the surface albedo [Lohila et al., 2010].

In this study, we investigated peatland dynamics since the early Holocene initiation, including the postdrainage phase. We used a multiple-core approach, and we reconstructed vegetation changes, lateral expansion, and C accumulation patterns. Detected changes were interpreted in the context of Holocene climate variations. We combined our palaeoecological data with contemporary greenhouse gas measurement data in order to reconstruct long-term radiative forcing through time.

2. Methods

2.1. Site

Kalevansuo peatland is located in southern Finland (60°38’49”N, 24°21’23”E; elevation 123 m above sea level; Figure 1a). The regional mean annual temperature and precipitation in 1971–2000 were 4.3°C and 647 mm, respectively [Drebs et al., 2002]. The predrainage vegetation was characteristic of a nutrient poor dwarf-shrub pine bog. In 1969 the bog was drained for forestry, resulting in a lowered water table down to ~40 cm below the peat surface, and in 1973 it was fertilized with phosphorus and potassium, following Finnish guidelines.

The peatland at Kalevansuo in southern Finland—examined in this study—is not in a natural state, as it has been drained for forestry. In general, drainage for forestry results in a lowering of the water table through ditching, and this causes rapid changes to vegetation, namely, an increased coverage of trees, forest mosses, and shrubs at the expense of sedges, mire shrubs, and Sphagnum species [Laine et al., 1995]. The lower water table generally results in subsidence of the peat surface and increased peat bulk density [Minkkinen and Laine, 1998]. Drainage affects the C balance by changing, for example, plant productivity, litter quality, decomposition rates, and the residence time of organic matter in aerated conditions [Laiho, 2006]. Typically, these changes result in a decrease in CH4 emissions [Nykänen et al., 1998]. However, whether drained peatlands function as C sinks or sources in the long term is dependent on peatland type, climate, and the extent of change in the water level [Ojanen et al., 2013].

Approximately 4% of the northern peatland area is estimated to have been drained for forestry [Gorham, 1991; Paavilainen and Pääväänen, 1995]. This type of peatland management is especially prevalent in Finland, where ~57,000 km² of peatlands were drained for forestry during the last century, equivalent to 55% of the total original peatland area [Turunen, 2008]. Here drainage has often increased the C uptake [Minkkinen and Laine, 1998; Minkkinen et al., 2002], and correspondingly, it has been estimated that the widespread drainage activities in Finland have had a cooling effect on the atmosphere [Minkkinen et al., 2002], although this effect is partly negated by changes in the surface albedo [Lohila et al., 2010].

In this study, we investigated peatland dynamics since the early Holocene initiation, including the postdrainage phase. We used a multiple-core approach, and we reconstructed vegetation changes, lateral expansion, and C accumulation patterns. Detected changes were interpreted in the context of Holocene climate variations. We combined our palaeoecological data with contemporary greenhouse gas measurement data in order to reconstruct long-term radiative forcing through time.

2. Methods

2.1. Site

Kalevansuo peatland is located in southern Finland (60°38’49”N, 24°21’23”E; elevation 123 m above sea level; Figure 1a). The regional mean annual temperature and precipitation in 1971–2000 were 4.3°C and 647 mm, respectively [Drebs et al., 2002]. The predrainage vegetation was characteristic of a nutrient poor dwarf-shrub pine bog. In 1969 the bog was drained for forestry, resulting in a lowered water table down to ~40 cm below the peat surface, and in 1973 it was fertilized with phosphorus and potassium, following Finnish guidelines for forestry on drained peatlands. Following the drainage in 1969, the growth of the tree stand increased, dominated by Pinus sylvestris L., with occasional Betula pubescens Roth and understory Picea abies L. Dwarf shrubs consist of Ledum palustre L., Vaccinium uliginosum L., V. vitis-idea L., V. myrtillus L., Empetrum nigrum L., and Calluna vulgaris (L.) Hull. Eriophorum vaginatum L. and Rubus chamaemorus L. occur in the field layer. The moss layer consists of forest mosses Pleurozium schreberi (Brid.) Mitt., Dicranum polysetum Sw., Aulacomnium palustre (Hedw.) Schwägr., and Polytrichum strictum (Br.) J. Schimp. Moist patches support peat mosses, such as Sphagnum angustifolium (Russ.) C. Jens., S. magellanicum Brid., and S. russowii Warnst.

Kalevansuo covers an area of ~0.9 km² and the peat depth ranges from 0.4 to 3 m (Table 1) [see Lohila et al., 2011], with the deepest parts found in the west, close to sampling points A and C (Figure 1b). In the northern and eastern parts of the peatland, north of point G and east of point Q (Figure 1b), the peat depth is less than...
0.5 m. The site has previously been studied in regard to greenhouse gas fluxes and the contemporary C balance [Badorek et al., 2011; Koskinen et al., 2014; Lohila et al., 2011; Ojanen et al., 2012; Pihlatie et al., 2010].

2.2. Sampling and Dating

In 2012, peat cores were collected by using a Russian peat corer with a 5 cm diameter cylinder from 19 sampling points that cover the entire peatland (Figure 1b). Of the collected cores, eight (A–H) were studied for vertical peatland development, while the other cores (I–T) were used to study lateral expansion.

![Figure 1.](image)

**Figure 1.** (a) Location of Kalevansuo (star symbol) in Finland and (b) location of the coring points (A–T). The coring-point-specific basal peat ages are shown. The gray coloring indicates estimated isochrones based on basal ages.

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth (cm)</th>
<th>Radiocarbon Age (year B.P. ±SD)</th>
<th>Calibrated Age (cal B.P.)</th>
<th>Modeled 95% Conf. Interval (cal B.P.)</th>
<th>Dated Material</th>
<th>Laboratory Code</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>194–196</td>
<td>7,660 ± 40</td>
<td>8,450</td>
<td>8,237–8,623</td>
<td>Sedge remains</td>
<td>Poz-67161</td>
</tr>
<tr>
<td>A</td>
<td>238–240</td>
<td>9,248 ± 62</td>
<td>10,420</td>
<td>10,131–10,721</td>
<td>Wood, Equisetum remains</td>
<td>Hela-3299</td>
</tr>
<tr>
<td>B</td>
<td>18–20</td>
<td>350 ± 30</td>
<td>395</td>
<td>310–499</td>
<td>Sphagnum stems</td>
<td>Poz-67165</td>
</tr>
<tr>
<td>B</td>
<td>146–148</td>
<td>8,510 ± 50</td>
<td>9,510</td>
<td>9,111–9,770</td>
<td>Sedges</td>
<td>Poz-67166</td>
</tr>
<tr>
<td>B</td>
<td>178–180</td>
<td>9,143 ± 72</td>
<td>10,325</td>
<td>10,247–10,903</td>
<td>Equisetum</td>
<td>Hela-3309</td>
</tr>
<tr>
<td>C</td>
<td>22–24</td>
<td>62 ± 30</td>
<td>95</td>
<td>16–263</td>
<td>Wood</td>
<td>Hela-3300</td>
</tr>
<tr>
<td>C</td>
<td>78–80</td>
<td>3,734 ± 40</td>
<td>4,085</td>
<td>3,816–4,348</td>
<td>Wood, moss stems</td>
<td>Hela-3301</td>
</tr>
<tr>
<td>C</td>
<td>130–132</td>
<td>7,038 ± 56</td>
<td>7,875</td>
<td>7,509–8,119</td>
<td>Wood</td>
<td>Hela-3302</td>
</tr>
<tr>
<td>C</td>
<td>142–144</td>
<td>8,070 ± 50</td>
<td>8,995</td>
<td>8,413–9,252</td>
<td>Moss stems</td>
<td>Poz-67162</td>
</tr>
<tr>
<td>C</td>
<td>158–160</td>
<td>8,564 ± 67</td>
<td>9,540</td>
<td>9,378–10,000</td>
<td>Ericaceae seeds</td>
<td>Hela-3303</td>
</tr>
<tr>
<td>D</td>
<td>46–48</td>
<td>565 ± 30</td>
<td>600</td>
<td>519–653</td>
<td>Wood, sedges</td>
<td>Poz-72587</td>
</tr>
<tr>
<td>D</td>
<td>112–113</td>
<td>5,433 ± 48</td>
<td>6,235</td>
<td>5,544–6,620</td>
<td>Wood, seeds, charcoal</td>
<td>Hela-3304</td>
</tr>
<tr>
<td>E</td>
<td>122–124</td>
<td>2,580 ± 30</td>
<td>2,735</td>
<td></td>
<td>Wood</td>
<td>Poz-67164</td>
</tr>
<tr>
<td>F</td>
<td>120–122</td>
<td>8,514 ± 66</td>
<td>9,505</td>
<td></td>
<td>Wood, seeds, Equisetum</td>
<td>Hela-3306</td>
</tr>
<tr>
<td>G</td>
<td>52–54</td>
<td>500 ± 30</td>
<td>525</td>
<td>454–630</td>
<td>Sphagnum stems</td>
<td>Poz-72588</td>
</tr>
<tr>
<td>G</td>
<td>80–82</td>
<td>3,102 ± 36</td>
<td>3,305</td>
<td>1,981–4,762</td>
<td>Wood</td>
<td>Hela-3307</td>
</tr>
<tr>
<td>H</td>
<td>40–42</td>
<td>2,093 ± 36</td>
<td>2,065</td>
<td></td>
<td>Wood, charcoal</td>
<td>Hela-3308</td>
</tr>
<tr>
<td>I</td>
<td>106–108</td>
<td>3,925 ± 42</td>
<td>4,360</td>
<td></td>
<td>Wood, charcoal</td>
<td>Hela-3310</td>
</tr>
<tr>
<td>J</td>
<td>33–35</td>
<td>2,014 ± 36</td>
<td>1,965</td>
<td></td>
<td>Sphagnum stems, charcoal</td>
<td>Hela-3311</td>
</tr>
<tr>
<td>K</td>
<td>88–90</td>
<td>4,788 ± 52</td>
<td>5,515</td>
<td></td>
<td>Wood, seeds</td>
<td>Hela-3312</td>
</tr>
<tr>
<td>L</td>
<td>143–145</td>
<td>5,257 ± 53</td>
<td>6,040</td>
<td></td>
<td>Wood</td>
<td>Hela-3313</td>
</tr>
<tr>
<td>M</td>
<td>211–213</td>
<td>9,281 ± 73</td>
<td>10,460</td>
<td></td>
<td>Wood</td>
<td>Hela-3314</td>
</tr>
<tr>
<td>N</td>
<td>160–162</td>
<td>7,560 ± 68</td>
<td>8,370</td>
<td></td>
<td>Wood, seeds</td>
<td>Hela-3315</td>
</tr>
<tr>
<td>P</td>
<td>60–62</td>
<td>4,567 ± 51</td>
<td>5,215</td>
<td></td>
<td>Wood</td>
<td>Hela-3316</td>
</tr>
<tr>
<td>Q</td>
<td>230–232</td>
<td>8,786 ± 68</td>
<td>9,820</td>
<td></td>
<td>Sphagnum stems, wood, seeds</td>
<td>Hela-3317</td>
</tr>
<tr>
<td>R</td>
<td>16–18</td>
<td>MODERN</td>
<td></td>
<td></td>
<td>Wood, seeds</td>
<td>Hela-3318</td>
</tr>
<tr>
<td>S</td>
<td>59–61</td>
<td>6,389 ± 59</td>
<td>7,325</td>
<td></td>
<td>Wood, Equisetum</td>
<td>Hela-3319</td>
</tr>
<tr>
<td>T</td>
<td>103–105</td>
<td>6,809 ± 59</td>
<td>7,650</td>
<td></td>
<td>Needles, wood, seeds</td>
<td>Hela-3320</td>
</tr>
</tbody>
</table>

**Table 1.** Dated Peat Samples From Kalevansuo

The table lists the dated peat samples with their core numbers, depths, radiocarbon ages, calibrated ages, modeled 95% confidence interval, dated materials, and laboratory codes.
Selected plant remains from 27 peat samples (2 cm thick slices) were radiocarbon dated by the accelerator mass spectrometry method at the Finnish Museum of Natural History (LUOMUS, Helsinki, Finland) and at Poznan Radiocarbon Laboratory (Poznan, Poland; Table 1). Basal peat age was determined from the bottom layer of all cores. Additional depths between the core bottom and top were dated from points A, B, C, D, and G. These dated levels were chosen based on detected major vegetation shifts. The $^{14}$C ages were calibrated using IntCal13, and the calibrated two-sigma median age was used as an age estimation [Reimer et al., 2013], expressed as calibrated years before present (cal B.P.; present = 1950). Age-depth modeling was performed for the cores with multiple dates, using BACON software [Blaauw and Christen, 2011], and the weighted mean age was used to estimate the age of intermediate peat layers.

**2.3. Macrofossil Analysis**

Plant macrofossil composition was analyzed for cores A–H. Two-centimeter peat slices, selected with an 8 cm depth interval, were analyzed. The analysis mainly followed the Quadrat and Leaf Count protocol described by Mauquoy and Van Geel [2007] and modified by Välimäki et al. [2007]. Specifically, volumetric 5 cm$^3$ subsamples were cleaned under running water by using a 140μm sieve. No chemical treatment was necessary. The material that remained on the sieve was first examined under a low-power stereomicroscope. Volumetric percentages were determined for different plant types, e.g., Sphagnum, Cyperaceae, and Ericaceae. Simultaneously, macroscopic charcoal particles were counted. If the proportion of mosses exceeded 10%, a high-power light microscope was used for species level identification. The macrofossil assemblages were clustered by using CONISS [Grimm, 1987] and divided into zones that corresponded to the main clusters.

**2.4. Carbon Accumulation**

Cores A–H were analyzed for bulk density (BD) and loss on ignition (LOI) from alternate peat slices. Carbon content of the dried peat, as a percentage of organic matter (% of OM by mass), was analyzed by using duplicate analyses with a Vario Max CNS elemental analyzer (Elementar Analysensysteme GmbH, Germany). Eight depths (26, 34, 74, 98, 158, 178, 206, and 226 cm) from core A were analyzed. The samples covered all the different successional stages. We then used these successional stage-specific C content values, together with LOI, to estimate the C content (as % of dry bulk peat mass) of cores A–H. The calibrated median age of dated samples, BD values, and C content of the peat were used to calculate the apparent C accumulation rate (CAR) of cores A–H.

**2.5. Holocene Carbon Dynamics**

We used peatland expansion, successional stage, and CAR data to reconstruct the CH$_4$ and CO$_2$ fluxes between the Kalevansuo peatland and the atmosphere during the Holocene. To this end, the total peatland area was divided in four nonoverlapping subareas based on basal dates (Figure 1b), in which lateral expansion was assumed to have occurred linearly between basal dates. Each subarea included data from at least one coring point A–H. Plant data derived from the A–H points were used to describe the mean successional development of the respective subarea. In order to estimate the Holocene CH$_4$ flux for each successional stage, we used previously measured contemporary flux data for Finnish peatlands: 11.2 ± 7.1 (±SD) g CH$_4$C m$^{-2}$yr$^{-1}$ for rich fens, 18.0 ± 9.7 g CH$_4$C m$^{-2}$yr$^{-1}$ for poor fens, and 3.7 ± 3.4 g CH$_4$C m$^{-2}$yr$^{-1}$ for ombrotrophic bogs [Minkkinen and Ojanen, 2013, and references therein]. We used the measured contemporary flux in Kalevansuo, −0.09 ± 0.15 g CH$_4$C m$^{-2}$yr$^{-1}$, for the drained bog stage [Lohila et al., 2011]. The CO$_2$ flux that affects the atmospheric CO$_2$ pool was derived from the C accumulation rate:

$$\text{FCO}_2(t) = - \text{CAR}(t) \quad (1)$$

where FCO$_2$ is the CO$_2$ flux (g CO$_2$C m$^{-2}$yr$^{-1}$) at time $t$ (cal B.P.) and CAR is the C accumulation rate (g C m$^{-2}$yr$^{-1}$). Here we followed the micrometeorological sign convention for CO$_2$ and CH$_4$ fluxes: a positive sign indicates a flux from the ecosystem to the atmosphere (emission), and a negative sign indicates a flux from the atmosphere to the ecosystem (uptake). Equation (1) involves an implicit assumption that the CH$_4$ emitted to the atmosphere is rapidly oxidized to CO$_2$ [Frolking et al., 2006]. Similarly, the C lost from the peatland as dissolved organic carbon (DOC) was not taken into account in the CO$_2$ flux, as we assumed that most of the DOC rapidly mineralizes to CO$_2$ and returns to the atmosphere [e.g., Evans et al., 2015; Köhler et al., 2002]. The CO$_2$ fluxes of multiple cores were averaged for subareas. In one core (C), the very high CAR values...
in the top ~30 cm were excluded, as these are clearly associated with young undecomposed material [Clymo, 1984] and are therefore not comparable with CAR values from the deeper peat. The CO₂ flux of the drained peatland stage was based on the measured contemporary soil CO₂ balance in Kalevansuo: −65.5 g CO₂-C m⁻² yr⁻¹ [Lohila et al., 2011; Ojans et al., 2012]. Finally, the average CO₂ and CH₄ fluxes (g C m⁻² yr⁻¹) per subarea were multiplied by the estimated peat area (m²) of that subarea. These CO₂ and CH₄ fluxes per subarea per time step were then summed to yield long-term peatland-scale flux reconstructions (g CO₂-C yr⁻¹ and g CH₄-C yr⁻¹, respectively) covering the Holocene period.

2.6. Radiative Forcing Model
The annual peatland-scale CO₂ and CH₄ fluxes reconstructed for Kalevansuo were used to calculate the apparent climate impact of the peatland development throughout the Holocene. This effect is expressed as radiative forcing (RadF), which quantifies the change in net irradiance at the top of the troposphere [Myhre et al., 2013]. We used a sustained pulse-response model [Lohila et al., 2010] to calculate the RadF that resulted from the changes in atmospheric mixing ratios of CO₂ and CH₄, due to CO₂ uptake and CH₄ emissions in Kalevansuo. Previously, this model has been used to reconstruct the Holocene RadF of a subarctic fen [Mathijssen et al., 2014], and a similar model has been applied to estimate the RadF due to C dynamics of northern peatlands [Frolking et al., 2006; Frolking and Roulet, 2007] and Siberian thermokarst lakes [Walter Anthony et al., 2014] during the Holocene.

In our model, the atmospheric mixing ratio change Δ忻 that generates RadF is calculated by integrating the response function related to an instantaneous (positive or negative) emission pulse over time, t. By describing the surface flux dynamics φ(t) by a sustained series of such pulses, we can write

\[ \Delta \chi(t) = \int_0^t \phi(t)f(t-t)dt \]  

(2)

where k denotes the emission-mixing ratio conversion factor resulting from the instantaneous and complete atmospheric mixing assumed in the model and f is the impulse response function that indicates the fraction of the change in mixing ratio due to an emission pulse remaining in the atmosphere. For CO₂ we use the parameterization of f that was derived from a multimodel ensemble of simulations with process-based Earth system models by Joos et al. [2013]. This impulse response consists of a weighted sum of four exponential functions that represent different perturbation time scales involved in the global biogeochemical CO₂ cycles. The longest time scale effectively corresponds to a permanent mixing ratio change with a weight of 22% of each CO₂ pulse [Joos et al., 2013]. For the CH₄ pulse, we assume a first-order decay with a perturbation time scale of 12 years [Myhre et al., 2013]. The impulse response functions are assumed not to vary during the calculation period, and thus, we ignore any potential effect of global changes in C cycling and atmospheric oxidation processes.

The RadF(t) resulting from the Δ忻(t) calculated with equation (2) is estimated with the formulae employed by the Intergovernmental Panel on Climate Change [Myhre et al., 2013, Table 8. SM.1]. These parameterized relationships, which are based on detailed radiation transfer models, provide the instantaneous RadF as a function of atmospheric mixing ratio as related to the unperturbed mixing ratio of each gas. Using these relationships, the RadF resulting from the long-term flux dynamics of Kalevansuo can be calculated as a marginal change in a fixed background mixing ratio [Lohila et al., 2010]. Thus, in addition to the annually varying peatland-scale CO₂ and CH₄ fluxes reconstructed for Kalevansuo, the input of our RadF model consists of annual background mixing ratios of CO₂ and CH₄. For these we assume constant values for the preindustrial era (until 1780) with measured increases included thereafter [Meinshausen et al., 2011; World Meteorological Organization, 2016].

3. Results
3.1. Lateral Peatland Expansion
The basal peat dates (Table 1) indicate that peat initiation in Kalevansuo started ~10,500 cal B.P. in the central western part of the current peatland (Figure 1b). By 8000 cal B.P., an elongated fen had formed, after which peatland expansion was directed toward the north, south, and west. The last and relatively rapid expansion spurt toward the north and east after 3000 cal B.P. was probably regulated by relatively steep slopes (up to 9°) that constrained expansion elsewhere in the basin.
Figure 2. Plant macrofossils (left plots), bulk density (BD; solid line), and loss on ignition (LOI; dashed line) from cores A–H. The proportions of grouped macrofossil assemblages are shown. The groups are fen species, *Eriophorum vaginatum*, bog mosses, shrubs and trees, forest mosses, and unidentified organic matter. Asterisks indicate the presence of charcoal as follows: Single asterisk indicates only small-sized (<1 mm) charcoal particles or less than 10 larger (>1 mm) particles per 5 cm³ sample. Double asterisk indicates many small particles and 10 to 20 large particles per 5 cm³ sample. Triple asterisk indicates many small particles and >20 large particles per 5 cm³. The inferred successional stages RF (rich fen), PF (poor fen), OB (ombrotrophic bog), and DB (drained bog) are divided by gray lines. The black arrows indicate dated levels and are labeled with the calibrated median age of the sample (see also Table 1).
3.2. Peatland Succession

The macrofossil records from cores A–H (Figure 2) show a distinct development in vegetation type, from an initial fen peat type toward a pine bog. Furthermore, the effect of drainage in 1969 is visible in the top layers of the samples. Four vegetation assemblage zones can be distinguished that define different developmental stages. These stages consist of rich fen, poor fen, ombrotrophic bog, and drained bog.

3.2.1. Rich Fen

The rich fen (RF) phase represented the initial stage and is typically characterized by a dominance of Cyperaceae. In cores C and F, the occurrence of aquatic species (Isoëtes, Nymphae sp., and Potamogeton sp.), accompanied by rich fen species (Trichophorum alpinum, Menyanthes trifoliata, Scorpidium scorpioides, and Sphagnum teres), indicates that peat initiation at this location occurred through terrestrialization. In contrast, limnic elements were not present in the other cores. Instead, the bottom peat layers were characterized by the presence of roots of Ericaceous shrubs and by birch (Betula sp.) woody remains. The unidentified organic matter peat fraction dominated the RF phase.

The start of RF was dated to ~10,500 cal B.P. for points A and B, while at points C and F fen peat accumulation started ~9500 cal B.P. The RF phase and peat accumulation started much later at points D and E: ~6230 and 2740 cal B.P., respectively (Figures 2 and 3 and Table 1). The RF phase lasted from 400 to 4500 years.

3.2.2. Poor Fen

The poor fen (PF) phase was dominated by E. vaginatum. All analyzed cores included a clear PF phase (Figure 2), although in core D the upper and lower boundaries were diffused since E. vaginatum codominated with other Cyperaceae or bog-mosses. Other species that typically coexisted with E. vaginatum were Cyperaceae, ericaceous shrubs, birch, and conifers. In points B and C, the PF phase started at 10,000–9000 cal B.P., in A and F at ~8000 cal B.P., and in D, E, G, and H between 3500 and 2000 cal B.P. (Figure 3). The PF phase lasted between 600 and 7500 years. The presence of macroscopic charcoal particles indicated frequent burning throughout the PF phase.

3.2.3. Ombrotrophic Bog

The ombrotrophic bog (OB) vegetation was dominated by Sphagnum species together with woody (Ericaceae) species. This phase was present in all studied points (Figure 2), although at point F, the OB layer was only 15 cm thick, while at other points this layer was between 25 and 85 cm thick. The start...
of the OB phase (i.e., the fen-bog transition) was dated to two separate time windows: ~4000–3500 cal B.P. (points A and C) and ~1000–500 cal B.P. (points B and D–H) (Figure 3). However, it should be noted that the ombrotrophication date for points E, F, and H was tentative as only basal dates were available.

Macrofossil records from several points showed variation in vegetation composition within the OB phase. At point D, the OB phase started with a combination of S. fuscum and S. magellanicum together with E. vaginatum, but halfway during the OB phase the amount of these species either decreased or they disappeared, and the amount of Sphagnum sect. cuspidata increased. At point B, the mixture of S. magellanicum/ Sphagnum sect. acutifolia/E. vaginatum was replaced by Sphagnum sect. cuspidata species at this time. These changes indicate an increase in peat surface moisture levels. At point C, the hummocky OB phase was interrupted by a series of fires, as evidenced by a 10 cm thick layer rich in charcoal particles (Figure 2). This period of frequent fires was dated between 1270 and 690 cal B.P. and was characterized by an increase in E. vaginatum abundance, whereas Sphagnum remains were absent (Figure 2). Postfire communities again included S. magellanicum and subsequently S. fuscum and Ericaceae. Large amounts of charcoal were also found at point A, concentrated around a peat depth of 35 cm. However, the species assemblages slightly differed from point C: no increase in abundance of E. vaginatum could be observed (Figure 2).

### 3.2.4. Drained Bog

In the drained bog (DB) phase, Sphagnum mosses were replaced by typical forest mosses, such as Pleurozium schreberi and Dicranum polysetum, which were accompanied by hummock brown mosses, e.g., Polytrichum strictum and Aulacomnium palustre. The amount of Ericaceae remains increased, and the presence of pine and birch became more continuous. In general, the DB phase covered the top peat layers (i.e., the upper 5 to 15 cm), although at points F and G the distinction between OB and DB was not very clear and at point H the DB phase seemed to be missing (Figure 2).

#### 3.3. Carbon Accumulation

Bulk density (BD) values ranged from 0.03 to 0.25 g cm\(^{-3}\) (Figure 2), with a mean of 0.09 g cm\(^{-3}\). BD decreased after ombrotrophication. Successional stage-specific mean BD (±SD) values were RF 0.11 ± 0.03, PF 0.07 ± 0.03, and DB 0.07 ± 0.02 g cm\(^{-3}\). LOI values showed more short-term variation than BD (Figure 2), and the zones dominated by Sphagnum species had the highest LOI values. The successional stage-specific LOI values were RF 86.8 ± 12%, PF 89.1 ± 9%, OB 96.1 ± 5%, and DB 91.9 ± 7%.

The measured C content (point A) ranged from 51 to 71% of total organic matter. Samples dominated by Sphagnum had a much lower C content: 52% (±1.9%); than those dominated by sedges (E. vaginatum or other Cyperaceae): 64% (±5.3%). The higher C content in sedge dominated samples was partly due to charcoal present in these samples. However, the level of charcoal presence was similar in the samples analyzed for C content as in all RF and PF macrofossil samples (charcoal present in 60% of samples), justifying the use of these high C content values to calculate CAR values during the RF and PF stages.

In general, C accumulation rates (CAR) ranged from 5.5 to 27.5 g C m\(^{-2}\) yr\(^{-1}\) (Figure 4). In cores A, B, and C, CAR decreased notably after the initial rich fen phase (Figure 4, see also Figure 2). Between 8000 and 600 cal B.P., CAR values ranged from 5.5 to 10.5 g C m\(^{-2}\) yr\(^{-1}\). After 600 cal B.P., CAR clearly increased in some of the cores, from 5–9 to 20–25 g C m\(^{-2}\) yr\(^{-1}\). Core C showed high CAR values of 81 g C m\(^{-2}\) yr\(^{-1}\) between 95 cal B.P. and the present (Figure 4).

#### 3.4. Holocene Carbon Dynamics

For the majority of the past 10,500 years, the magnitude of the reconstructed CH\(_4\) and CO\(_2\) fluxes at Kalevansuo (Figure 5b) increased reflecting an expanding peatland area (Figure 5a). The CH\(_4\) flux increased from zero to \(13 \times 10^6\) g CH\(_4\)-C yr\(^{-1}\) and the CO\(_2\) flux from zero to \(-8 \times 10^6\) g CO\(_2\)-C yr\(^{-1}\). However, some slight fluctuations were also visible in this period (Figure 5b), associated with successional changes in various parts of the peatland (Figure 3). By 600 cal B.P., the whole peatland area was covered by bog vegetation (Figure 3) and this is reflected as a large increase in CO\(_2\) uptake, from \(-8\) to \(-20 \times 10^6\) g CO\(_2\)-C yr\(^{-1}\), and a decrease in CH\(_4\) emissions, from 11 to \(3 \times 10^6\) g CH\(_4\)-C yr\(^{-1}\). After drainage, CO\(_2\) uptake further increased to \(-58 \times 10^6\) g C yr\(^{-1}\) and CH\(_4\) fluxes were reduced to \(-0.08 \times 10^6\) g CH\(_4\)-C yr\(^{-1}\). As a result of these long-term C fluxes, the amount of C currently stored in Kalevansuo is estimated at \(44 \times 10^9\) g C, i.e., 49 kg C m\(^{-2}\) on average (Figure 5b).
3.5. Radiative Forcing (RadF)

The apparent RadF of Kalevansuo was positive (warming impact) during the first ~7000 years following peatland initiation (Figure 6). The maximum RadF (0.7 × 10⁻⁸ W m⁻²) occurred around 7000 cal B.P. and then decreased because of the sustained C accumulation. The RadF became negative (cooling impact) after 4000 cal B.P., but around 2000 cal B.P. it returned toward positive values again for a few hundred years. This rise in RadF was associated with the expansion of the poor fen peat area (Figure 3), and the corresponding

![Figure 4](image.png)

Figure 4. Carbon accumulation rate (CAR; g C m⁻² yr⁻¹) in cores A–H.

![Figure 5](image.png)

Figure 5. Reconstructed (a) peat area and (b) C dynamics integrated over the total peatland area, CO₂ and CH₄ fluxes (10⁶ g CO₂-C yr⁻¹; 10⁶ g CH₄-C yr⁻¹), and cumulative C stored (10⁹ g C) in Kalevansuo over time.
increase in CH$_4$ emissions. The rapid change in both CO$_2$ and CH$_4$ fluxes after 600 cal B.P. (Figure 5b) resulted in a rapid decrease in the magnitude of the corresponding RadF components (Figure 6). This change was driven by ombrotrophication of the peatland. The current RadF of Kalevansuo is estimated at $-7.8 \times 10^{-8}$ W m$^{-2}$, taking into account its entire C dynamics history, which indicates a consistent atmospheric cooling effect due to sustained CO$_2$ uptake.

4. Discussion

4.1. Peatland Development and Driving Factors

Peat accumulation at Kalevansuo began at a few individual locations from approximately 10,500 cal B.P. By 9000 cal B.P., several initiation loci had merged and formed an elongated fen (Figure 1b). Early Holocene peatland initiation and expansion at Kalevansuo corresponds with the reported peak in Fennoscandian peatland initiation at 11,000–9000 cal B.P. [Ruppel et al., 2013], which was enabled by early Holocene regional climate with warm summers [Väliranta et al., 2015]. By ~8500 cal B.P., the initial rich fen was replaced by poor fen, dominated by E. vaginatum (Figure 3). This phase commonly precedes ombrotrophication [Hughes and Barber, 2003; Tuittila et al., 2013; Väliranta et al., 2016]. By 2500 years (following initiation) the peatland had covered an area of 0.15 km$^2$ (Figure 5a).

Between 8000 and 5000 cal B.P., the peatland area increased to ~0.40 km$^2$ (Figure 5a). The vegetation composition was characteristic for the rich fen type, although the older center of the peatland remained as poor fen (Figure 3). In some cases, variation in microclimate conditions can be seen directly as shifts in peatland vegetation communities [e.g., Tuittila et al., 2007; Väliranta et al., 2007]. In Kalevansuo, the poor fen phase that started ~9000 cal B.P. and was dominated by E. vaginatum may be linked to climate changes that induced lower atmospheric humidity levels [Seppä et al., 2005; Väliranta et al., 2005]. Eriophorum vaginatum is known to tolerate and thrive under variable and low water table levels [Kummerow et al., 1988]. The change in vegetation composition at 9000 cal B.P. was accompanied by decreasing CAR values (Figure 4), but no change in peatland expansion was observed (Figure 1b).

Between 5000 and 3000 cal B.P., the peatland expanded until the steep slopes to the south, west, and north-west constrained further lateral expansion (Figure 1b). The peatland area increased to 0.60 km$^2$ (Figure 5a). In contrast to the previous periods, the new peatland area was inhabited by E. vaginatum rather than rich fen species. The western part of Kalevansuo transformed into a bog, while the other parts remained as poor fen until 3000 cal B.P. A small part also remained as rich fen (Figure 3).

During the last 3000 years, Kalevansuo reached its current area of ~0.90 km$^2$ (Figure 5a). Between 3000 and 2000 cal B.P., the peatland expanded toward the north, and only recently toward the east (Figure 1b). Some of this newly formed area went through a short-lasting rich fen phase; however, by approximately 2000 cal B.P. these areas were covered by poor fen. Subsequently, between 1000 and 500 cal B.P., the remaining peatland area was transformed into bog (Figure 3). The total peatland area increased almost linearly through time (Figure 5a), and this includes the last 3000 years when expansion was partly constrained. In fact, the lateral expansion toward the north and east greatly accelerated during the last 3000 years (Figure 1b). This acceleration in expansion rate was not linked to a transition in the successional stages, as ombrotrophication did not occur for another 1000 years throughout most of the peatland (Figure 2). Instead, it may be that lateral expansion was driven by climate change toward cooler and wetter surface conditions.
conditions [Snowball et al., 2004]. This link between climate variation and lateral expansion has been previously observed by Korhola [1994], Turunen and Turunen [2003], and Ruppel et al. [2013]. Most occurrences of ombrotrophication in Kalevansuo occurred during the so-called Medieval Climate Anomaly (1000–550 cal B.P.; Figure 2), a warm and humid period in northern Europe [Diaz et al., 2011]. Moreover, ombrotrophication was not synchronous over the whole peatland (Figure 3), which suggests that this process was not totally driven by regional climate. In Kalevansuo, some vegetation assemblage reconstructions (points A and C; Figure 2) suggest that occasionally the bog development started when deep water tables prevailed locally between 5000 and 4000 cal B.P., whereas changes toward rising water tables have also been reported for bogs in southern Finland for this time period [Tuittila et al., 2007; Väilänta et al., 2007]. These discrepancies within relatively close distances make interpretation of climatic factors a challenge. Furthermore, peak core data from Kalevansuo indicate within-site variation in microtopographical conditions: a shift from hummock to hollow microtopography between ~400 and 200 cal B.P. in sample points B and D (indicated by a shift from Sphagnum sect. acutifolia species to Sphagnum sect. cuspidata species; Figures S2 and S4 in the supporting information) was not observed elsewhere in the peatland.

Fires may affect the vegetation development [Hughes and Barber, 2003; Sillasoo et al., 2011; Tuittila et al., 2007] and the C accumulation rates of peatlands [Pitkänen et al., 1999]. The studied peat cores from Kalevansuo showed evidence of frequent fires throughout the history of the peatland (Figure 2). Charcoal was repeatedly present in the basal peat layer in cores collected from points A, B, D, G, and H, and it is likely that fires played a facilitating role in peat initiation [cf. Tuittila et al., 2007] or areal expansion through paludification [Heinselm, 1975; Patterson et al., 1987]. At all sampling points, fires occurred frequently throughout the poor fen phase (Figure 2). The presence of woody species (Figures S1–S8) likely had a positive influence on fire frequency by providing fuel [Sillasoo et al., 2011]. Previous studies have similarly observed charcoal in peatland development phases dominated by E. vaginatum [Hughes et al., 2000; Mathijssen et al., 2016; Tuittila et al., 2007]. Tuittila et al. [2007] observed that fire disturbances repeatedly caused reversals in vegetation succession from Sphagnum spp. to E. vaginatum dominance. Hence, it seems plausible that the frequent fires in Kalevansuo maintained the poor fen phase with abundant Eriophorum cover and that the end of poor fen phase and the start of the bog phase might be related to decreasing fire disturbance. Large amounts of charcoal at ~35 cm depth from cores A and C and less distinct charcoal layers in the upper layers of the other cores (Figure 2; except core H) could be tentatively dated to the time window 1500 to 600 cal B.P. based on stratigraphical comparison with dated peat cores B–D. The timing of this charcoal layer corresponds with a period of elevated fire frequency (1000–500 cal B.P.) in three other bogs located in southern Finland and northern Estonia [Morris et al., 2014; see Mathijssen et al., 2016]. The observed elevated fire frequency between ~8000 and 4000 cal B.P. (cores A, B, C, D, and F; Figure 2) is similar to the results from peatlands in eastern Finland [Pitkänen et al., 1999]. Fire frequency may reflect climate conditions [Sillasoo et al., 2011], and it seems plausible that the period of high fire frequency between 8000 and 4000 cal B.P. can be attributed to regionally warm and dry climate conditions during the mid-Holocene [Seppä et al., 2005; Väilänta et al., 2005].

Our multiple-core approach indicates that single-core variations in vegetation should not be linked directly to regional climate but that they may merely reflect changes in local autogenic or indirect allogenic factors, such as fires.

4.2. Carbon Accumulation

The C content of peat samples from Kalevansuo (ranging from 51 to 68% of organic matter, or 54 to 71% where charcoal was present) was high in comparison to assembled northern peatland data (49.2 ± 2.4%) [Loisel et al., 2004]. The C content of Sphagnum-dominated samples in Kalevansuo (52 ± 1.9%) is at the upper limit of the range reported by Loisel et al. [2014] for this peat type, but somewhat lower than Sphagnum-dominated peat in Finland (55 ± 4.1%) [Minkkinen and Laine, 1998], and only slightly higher than the C content of Sphagnum peat in western Siberia (51 ± 0.005%) [Beilman et al., 2009]. In general, the C content of herbaceous and woody peat is higher than Sphagnum peat [Beilman et al., 2009; Loisel et al., 2014], but does not usually exceed 60%, although exceptions do exist [Chapman et al., 2009].

Bulk density (BD) in the ombrotrophic bog (0.07 ± 0.03 g cm⁻³) and drained bog (0.07 ± 0.02 g cm⁻³) were lower in comparison to the rich and poor fen peat samples (both 0.11 ± 0.03 g cm⁻³), while LOI showed a
less clear distinction between these successional stages (Figure 2). The BD and LOI values of the present study are similar to the values reported for northern peatlands by Loisel et al. (2014): BD of 0.076 ± 0.038 and 0.118 ± 0.075 g cm\(^{-3}\) for Sphagnum peat and herbaceous peat, respectively; LOI values of 94.3 ± 9.3% and 85.6 ± 15.4% for Sphagnum peat and herbaceous peat, respectively. Mäkilä and Laine (1998) reported comparable average BD values of Finnish sedge pine fens (0.13 ± 0.02 g cm\(^{-3}\)), accompanied with slightly higher LOI values (96.4 ± 1.8%) than Kalevansuo. The similarity of BD values from Kalevansuo with average values of undrained sites, and small BD variations, suggests that peat compaction caused by peat drainage played only a minor role.

The CAR values for Kalevansuo showed a pattern that can be linked to the successional stages (Figure 4): high values during the rich fen stage, low values during the poor stage, and an increase in values during the recent bog stages. However, this increase during the late part of the bog stage may be explained by the presence of less decomposed younger peat [Clymo, 1984]. This is supported by the fact that the two cores (A and C) that exhibited the transition to the bog phase earlier than the other cores (i.e., at ~4000 cal B.P.) (Figure 2) did not show elevated CAR values during the bog phase, but instead showed a slight decrease (Figure 4). The observed successional pattern in CAR values is similar to those found by Mäkilä [1997], Mäkilä et al. [2001], Mäkilä and Moisanen [2007], Peteet et al. [2016], and Mathijssen et al. [2016]. For the majority of the Holocene (i.e., 8000 to 600 cal B.P., which mainly corresponds to the poor fen phase in most of the cores), CAR values were lower (5.5 to 10.5 g C m\(^{-2}\) yr\(^{-1}\)) than the average values reported for poor fens in southern Finland (17 ± 8 g C m\(^{-2}\) yr\(^{-1}\) [Turunen et al., 2002]. These southern Finnish CAR values are, in turn, lower than the global northern peatland average (23 ± 2 g C m\(^{-2}\) yr\(^{-1}\)) [Loisel et al., 2014], but similar to those from continental western Canada (19.4 g C m\(^{-2}\) yr\(^{-1}\)) [Vitt et al., 2000]. Holocene CAR values from western Siberia ranged from 5 to 36 g C m\(^{-2}\) yr\(^{-1}\) [Beilman et al., 2009] and show the large variation between peatlands in the same region. Low peat and apparent C accumulation rates might be related to frequent fire events [see Mäkilä, 1997; Mathijssen et al., 2016], because a single fire event and associated peat burning may result in a loss of up to 2–4 kg C m\(^{-2}\) [Kuhry, 1994; Pitkänen et al., 1999]. In contrast, no evidence of peat loss through fire was observed by Mäkilä and Moisanen [2007] to explain low accumulation rates.

The estimated current C store in Kalevansuo of 49 kg C m\(^{-2}\) (Figure 5b) is low, when compared to the Siikaneva peatland complex, which has an estimated mean C store of 81 kg C m\(^{-2}\) [Mathijssen et al., 2016] and mean depth of 3.25 m [Tolonen et al., 1979]. It is, however, very close to the average C stock of pristine peatlands in the raised bog zone in Finland, 48 kg C m\(^{-2}\) [Turunen et al., 2002]. As the mean peat C% and bulk densities in pristine peatlands vary relatively little, peat C stock in a peatland depends mostly on peat thickness. In Finnish peatlands the mean peat thickness is about 1.1 m [Turunen et al., 2002], close to the average peat depth in Kalevansuo of 1.2 m. The latest global northern peatland C storage estimates, 118–155 kg C m\(^{-2}\) [Yu et al., 2010], would indicate much thicker peat deposits than measured in Kalevansuo or in Finland in general.

### 4.3. Radiative Forcing

The initial apparent net radiative forcing (RadF) of Kalevansuo was positive (warming), until it started to decrease after several thousand years, due to the long-term effect of CO\(_2\) removal from the atmosphere [Frolking et al., 2006]. Nearly 7000 years after peat initiation the net RadF became negative (cooling) (Figure 6); however, the short-term increase in reconstructed RadF at 1900 cal B.P. illustrates the RadF sensitivity to enhanced CH\(_4\) emissions due to increasing peatland area and peat type. The duration of the initial period with positive RadF is much longer than previously estimated for northern peatlands, i.e., 300–3500 years, using a RadF modeling approach similar to our study [Frolking and Roulet, 2007]. However, the long positive period is consistent with such model simulations, assuming CO\(_2\):CH\(_4\) flux ratios of 0.5 to 1, resulting in 5500–10,000 years of positive RadF [Frolking et al., 2006]. In Kalevansuo, the CO\(_2\):CH\(_4\) flux ratio was on average 0.7 between initiation and 600 cal B.P. (Figure 5b). Mathijssen et al. [2014] reported a period of positive RadF of 300–1700 years for a northern Finnish fen. However, their calculations failed to take into account the fact that the C emitted as CH\(_4\) eventually returns to the atmospheric CO\(_2\) pool. In contrast, Mathijssen et al. [2014] assumed that all C emitted as CH\(_4\) would be permanently removed from the atmosphere. This lead to an overestimation of CO\(_2\) removal from the atmosphere and an underestimation of the length of the period of positive RadF. If the current calculation method was applied to the data...
from Mathijsen et al. [2014], a switch from a positive to negative RadF would possibly disappear [Mathijsen, 2016].

The implicit assumptions of the RadF reconstruction are a source of uncertainties that cannot be fully quantified within the scope of this study. The development of site-specific RadF relies heavily on the estimated changes in peat area, which largely determines the estimates of the total carbon fluxes. This study makes use of 19 basal dates from a peatland of 0.9 km², which results in a relatively well constrained estimation of past peatland area when compared with, e.g., Korhola [1994], Mäkilä [1997], and Bauer et al. [2003]. The basal date density in Kalevansuo was similar to the one applied by Mäkilä and Moisanen [2007]. Nevertheless, it remains possible that the reconstructed peat area development is inaccurate by several hundred years.

In the reconstruction of CH₄ fluxes we used contemporary annual flux data from boreal peatlands. These data integrate spatial and seasonal variations in CH₄ emissions, but the current reconstruction of Holocene fluxes assumes that average flux rates have remained stable during this time. This is likely an unrealistic assumption as CH₄ emissions strongly respond to changes in temperature [e.g., Bellisario et al., 1999]. Consequently, past variations in temperature would have altered the mean CH₄ flux rates of peatland types. Incorporating the uncertainty of contemporary flux data, using the standard deviation as flux uncertainty, results in a maximum uncertainty range of 2.4 × 10⁻⁸ W m⁻² in RadF (data not shown) but does not affect the conclusions that can be drawn from our RadF calculations.

Compared to the uncertainties in RadF associated with estimations of peat area development and reconstruction of CH₄ fluxes, the applied method of equating apparent CAR values with permanent CO₂ uptake adds the largest uncertainty. This method implicitly underestimates the true CO₂ uptake during the Holocene, by assuming that any loss of C to the atmosphere occurs simultaneously with CO₂ uptake. This might not be a realistic representation of the past, since modeling results suggest that events, such as changes in water table or fires, may affect the apparent CAR values of peat layers accumulated up to 2000 years before the actual event occurred [Frolking et al., 2014]. Consequently, the applied apparent CAR values neglect the temporary storage of CO₂ buried in peat before the peat becomes decomposed or is combusted by fire, but it is unknown how large an effect this has on long-term RadF(CO₂).

### 4.4. Effect of Drainage

Drainage of Kalevansuo in 1969 lowered the water level to an average 40 cm below the peat surface [Lohila et al., 2011]. The resulting increase in peat aeration caused a rapid change in vegetation toward forest species (Figure 2). Following drainage for forestry, the changed moisture conditions and competition for nutrients and light may result in a complete transformation in vegetation within 30 to 50 years [Laine et al., 1995; Minkkinen et al., 1999; Tahvanainen, 2011]. Furthermore, drainage influences the C balance of a peatland as it is linked to changes in vegetation and decomposition processes [Laiho, 2006; Laiho and Laine, 1997; Minkkinen and Laine, 1998; Minkkinen et al., 1999]. However, this change in the C balance was not observed in our CAR values (Figure 4), because the chronological control and sampling resolution were not high enough for the topmost layers of the peat cores. However, if we extrapolate the contemporary C balance over the postdrainage time period, it results in a decrease of the RadF of the CH₄ flux to zero within 35 years after drainage (Figure 6). The cooling impact, i.e., decrease in RadF of CO₂, is not visible within the 40-year postdrainage time window (Figure 6), since it responds on longer time scales [Laine et al., 1996]. The RadF of CO₂ increased (became less negative) through the last century as the atmospheric concentrations of both CH₄ and CO₂ rose substantially from preindustrial values [Forster et al., 2007], thereby decreasing the radiative efficiency of further greenhouse gas additions. Kalevansuo has continued to sequester C even after drainage as a consequence of moderate water level drawdown [Laiho, 2006; Lohila et al., 2011; Minkkinen et al., 2002]. Under different postdrainage RadF scenarios with varying C dynamics, RadF should further decrease for 100–150 years after drainage and then after several centuries return to the predrainage level [Laine et al., 1996]. If the C that is sequestered in the tree biomass is taken into account, the further evolution of RadF would be dependent on the management processes implemented, such as different thinning practices [Laine et al., 1996; Minkkinen et al., 2002]. Land use changes also affect the surface albedo of boreal peatlands, and it has been shown that the magnitude of albedo-induced RadF can be similar to that caused by C sequestration [Lohila et al., 2010].
5. Conclusion

We traced peatland succession and the C dynamics of a drained bog at Kalevansuo through the Holocene period. The whole peatland went through the same succession phases from rich fen to poor fen to bog, but the timing of transitions varied between different parts of the peatland. In addition to autogenic succession, the vegetation development and lateral expansion were influenced by fires and changing regional climatic conditions. Drainage for forestry resulted in a replacement of typical bog mires (Sphagnum) by forest mires within 40 years of drainage. Carbon accumulation rates correlated with successional changes from rich to poor fen, but for the majority of the Holocene, accumulation rates were lower than the average rate of northern peatlands. We assume that C accumulation was limited, because of the C loss through frequent burning. Furthermore, variations between coring locations suggest that vegetation development data from a single core should not be directly linked to climatic variation but might reflect local changes unrelated to broad-scale climate conditions. The reconstructed RadF due to the atmosphere-peatland exchange of CO2 and CH4 was positive (warming effect) for the first ~7000 years of peatland development, after which it turned negative (cooling). RadF further decreased as a result of expansion of bog vegetation and later due to drainage. However, an expansion pulse of fen peatland around 2000 cal B.P. generated a temporary RadF increase. Our results emphasize the sensitivity of the peatland C balance to changes in peat type and areal coverage, which are in turn partly driven by regional climate.

Acknowledgments

We are grateful to Henriikka Kivilä for sampling assistance, to David Wilson for revising the English text, and to two anonymous reviewers for their helpful comments. We are grateful for financial support received from the University of Helsinki and ICOS-Finland funding received from the Finnish Ministry of Transport and Communication. The data used in this study are included in the table, figures, supplementary figures, or are available in the cited references.

References


