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CARBON DIOXIDE EXCHANGE ON CULTIVATED  
AND AFFORESTED BOREAL PEATLANDS

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## Abstract

Drainage and agricultural management practices fundamentally alter the moisture and nutrient conditions in peat soils, leading to high carbon dioxide (CO<sub>2</sub>) losses. Uncertainties in the emission estimates of CO<sub>2</sub> from cultivated peatlands are high, and the effects of different management practices on the carbon fluxes are not well understood. One possible land-use option for peat fields removed from agricultural use is afforestation, the effect of which on ecosystem CO<sub>2</sub> balance is practically unknown. The aim of this work was to improve the understanding of the biosphere-atmosphere exchange and annual balances of CO<sub>2</sub> on agriculturally-managed boreal peatlands. Micrometeorological eddy covariance measurements of CO<sub>2</sub> exchange were conducted above two drained peat soil sites with an agricultural history: a cultivated peat soil growing spring barley or forage grass and a 30-year old Scots pine plantation established on organic-soil cropland. Respiration fluxes at both sites were measured with the chamber technique. Partition of total ecosystem respiration ( $R_{tot}$ ) into autotrophic and heterotrophic components revealed that the soil respiration in the cultivated peat field was two to three times as high as that in cultivated clay and sandy soils. In addition to temperature, photosynthesis was a central factor controlling the dynamics of  $R_{tot}$ , and should be included in respiration models. While the wintertime emissions of CO<sub>2</sub> from the cultivated peatland were mainly controlled by the soil temperature, temporary deviations from this pattern were observed, especially during snow and soil thawing. At or immediately after the snow melt, decreased CO<sub>2</sub> efflux rates were observed, probably because of a reduction in gas diffusion rates in wet soil. In winter, significantly different temperature responses in CO<sub>2</sub> efflux were not found in bare and grass-covered soil. The number of snow-free days, as well as soil and air temperatures, seemed to control the year-to-year variation in winter CO<sub>2</sub> emission. The minimum and maximum values of daily net ecosystem exchange (NEE) during the growing season were rather similar for both of the studied ecosystems and for both crops, although slightly higher daily net CO<sub>2</sub> emission and lower net CO<sub>2</sub> uptake were observed in the peat soil growing grass. However, large differences in annual balances were observed. These were more attributable to the length of the season with net carbon uptake than to the magnitude of soil respiration. This length was controlled by the management of the ecosystems, being shorter with the annual crop than with the perennial crop, and being longest with the afforested peat field. All of the studied ecosystems acted as sources of CO<sub>2</sub> on an annual basis, NEE equalling 770 and 290 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup> in the barley and grass fields, respectively, and 50 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup> in the afforested organic-soil cropland. When taking into account the carbon in harvested biomass, the total annual carbon loss (excluding leaching and airborne erosion) in the barley and grass fields was about 340 and 450 g C m<sup>-2</sup> yr<sup>-1</sup>, respectively. Despite the much smaller annual NEE at the afforested site, the results indicate that even this ecosystem, and particularly the peat soil there, is losing carbon. This study suggests that the CO<sub>2</sub> loss from agricultural peat soils may be decreased with perennial crops or afforestation. These management strategies do not, however, turn these ecosystems into sinks of CO<sub>2</sub> as the carbon loss from decomposition of the old peat continues.

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## Tiivistelmä

Turvemaiden kuivatus ja maatalouskäyttö muuttavat maan kosteus- ja ravinneoloja, minkä seurauksena turpeesta vapautuu suuria määriä hiilidioksidia (CO<sub>2</sub>). Turvemaiden kuivatuksen ja viljelyn vaikutuksista CO<sub>2</sub>-päästöihin on kuitenkin olemassa vain vähän kvantitatiivista tietoa. Eräs mahdollinen maankäyttömuoto käytöstä poistetuille turvepelloille on metsitys, jonka seurauksia ekosysteemin hiilitaseen kannalta ei turvemaidella tunneta riittävästi. Tämän työn tarkoituksena on lisätä tietämystä ekosysteemin ja ilmakehän välisestä CO<sub>2</sub>-vaihdosta ja -taseista boreaalissa viljelyillä ja käytöstä poistetuilla turvepelloilla. CO<sub>2</sub>-vaihtoa mitattiin mikrometeorologisella kovarianssimenetelmällä kahdella ojitetulla turvemaidella. Toinen paikoista oli pelto, jolla kasvoi ensimmäisenä vuonna ohraa ja toisena nurmea. Toisella paikoista maatalouskäyttö oli loppunut 30 vuotta aiemmin, ja pelto oli metsitetty mänyllä. Respiraatiovoita mitattiin kammionmenetelmällä. Kokonaisrespiraatiosta mallin avulla erotettu maarespiraatio oli turvepelloilla 2–3 kertaa niin suurta kuin savi- ja hietapelloilla. Lämpötilan ohella yhteytysnopeus oli keskeinen respiraatiovoion selittäjä ja tulisi jatkossa sisällyttää vastaaviin malleihin. Vaikka turvepellon talviaikainen CO<sub>2</sub>-päästö selittyi parhaiten maan lämpötilalla, lumen ja maan sulaminen aiheuttivat hetkellisiä poikkeamia tästä riippuvuudesta siten, että lumen sulamisen aikaan ja heti sen jälkeen CO<sub>2</sub>:a vapautui maasta vähemmän kuin lämpötilan perusteella voitiin ennustaa. Tämä oli todennäköisesti seurausta diffuusion perustuvan kaasun kulkeutumisen hidastumisesta määssä maassa. Talvisessa CO<sub>2</sub>-päästöissä ei havaittu merkitsevää eroa kynnetyin ohrapellon ja nurmipellon välillä. Koko talvikauden päästö kolmelta talvelta korreloi positiivisesti lumettomien päivien lukumäärän sekä maan ja ilman lämpötilojen kanssa. Kasvukaudella päivittäisen CO<sub>2</sub>-vaihdon suuruus oli samaa luokkaa molemmissa ekosysteemeissä ja molemmilla peltokasveilla, vaikka sekä ekosysteemin kokonaisrespiraatio että yhteytys olivat hiukan pienempiä nurmipelloilla. Tästä huolimatta CO<sub>2</sub>:n vuositasessa havaittiin suuria eroja paikkojen ja vuosien välillä. Maarespiraatioissa ei viljelyllä ja metsitetyn turvemaan välillä havaittu merkitsevää eroa, vaan erot vuositasessa selittyivät enimmäkseen niiden päivien lukumäärällä, jona ekosysteemi toimi hiilidioksidin nieluna. Viljelytoimenpiteet vaikuttivat suuresti tämän jakson pituuteen: ohrapelto toimi lyhimmän jakson CO<sub>2</sub>-nieluna, kun taas metsitetyillä turvepelloilla tämä jakso oli pisin. Molemmilla paikoilla ja molemmilla viljelykasveilla CO<sub>2</sub>:n vuositaset olivat positiivisia. Ohra- ja nurmipellon vuosipäästöt olivat 770 ja 290 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup>, ja metsitetyn turvepellon päästö oli 50 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup>. Kun pelloilta korjatun sadon määrä huomioitiin, oli hiilen vuotuinen hävikki (lukuun ottamatta huuhtoutumaa ja tuulieroosiota) ohrapelloilta 340 ja nurmipelloilta 450 g C m<sup>-2</sup> yr<sup>-1</sup>. Pienemmästä vuosipäästöstä huolimatta näyttää siltä että myös metsitetyn pellon turvekerros vapauttaa CO<sub>2</sub>:a. Tämän tutkimuksen tulokset osoittavat, että monivuotisten kasvilajien käytöllä ja metsityksellä voidaan pienentää turvepeltojen CO<sub>2</sub>-päästöä, mutta näillä menetelmillä ei kyseisiä ekosysteemejä saada muutettua CO<sub>2</sub>:n nieluiksi. Täten hiilen hävikki tuhansia vuosia vanhasta turpeesta jatkuu edelleen.

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Helsinki, May 2008

Annalea Lohila

## DEFINITIONS

*(terms in italics have also been used in the text)*

**Afforestation** refers to establishment of trees on land which has long been without forest.

**Cultivation** refers to the growing of annual or perennial plant species for e.g. food or cattle feed. Includes activities like sowing, application of fertilizers, harvesting and ploughing the field.

**Mire** is peatland where peat is currently being formed and which is undrained and unmanaged. Synonym for *natural peatland* and *natural peat soils*.

**Peat** is partially decomposed plant material that has been accumulated in wetlands.

**Peat field** is a peatland with present or former agricultural activity. Synonym for *cultivated peat field* and *agricultural peat soil* and *organic-soil cropland*.

**Peatland** is a type of ecosystem with or without vegetation where organic matter has been accumulated as peat originating from decaying plant material. May be drained or undrained. The terms *peat soil* and *organic soil* may also refer to a peatland ecosystem.

**Plant respiration** is a process where the energy needed for the metabolic work of the plant is obtained from organic molecules (sugars) in plant cells, and which releases CO<sub>2</sub> and water. Synonym for *autotrophic respiration* and *autotrophic plant respiration*

**Soil respiration** is a process where soil microbes obtain energy and, by decomposing soil organic matter, release CO<sub>2</sub> as a byproduct. Synonym for *heterotrophic respiration* and *soil heterotrophic respiration*.

**Soil CO<sub>2</sub> efflux** means the flux of CO<sub>2</sub> from the soil into the atmosphere. May originate from microbial sources only or also include autotrophic sources such as roots.

**Wetland** is an ecosystem at the interface between a terrestrial and an aquatic ecosystem. Is at least some of the time water-saturated. Includes mires, but not drained peatlands where the water-table is lowered by ditching.

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## List of original publications

This thesis is based on the following four publications:

### Paper I

Lohila A., Aurela M., Regina K. and Laurila T. 2003. Soil and total ecosystem respiration in agricultural fields: effect of soil and crop type. *Plant and Soil*, 251, 303–317.

### Paper II

Lohila A., Aurela M., Regina K., Tuovinen J.-P. and Laurila T. 2007. Wintertime CO<sub>2</sub> exchange in a boreal agricultural peat soil. *Tellus*, 59B, 860–873.

### Paper III

Lohila A., Aurela M., Tuovinen J.-P. and Laurila T. 2004. Annual CO<sub>2</sub> exchange of a peat field growing spring barley or perennial forage grass. *Journal of Geophysical Research*, 109, D18116, doi:10.1029/2004JD004715.

### Paper IV

Lohila A., Laurila T., Aro L., Aurela M., Tuovinen J.-P., Laine J., Kolari P. and Minkkinen K. 2007. Carbon dioxide exchange above a 30-year-old Scots pine plantation established on organic-soil cropland. *Boreal Environment Research*, 12, 141–157.

## Summary of papers

**Paper I** presents chamber measurements on agricultural fields at Jokioinen. An empirical response model was constructed, which enabled the separation of plant and soil respiration components. Soil respiration components in different soil types were compared.

**Paper II** reports CO<sub>2</sub> flux data from three winters in 2000–2003 for the Jokioinen peat field measured by the eddy covariance method. The paper also includes chamber flux and soil concentration data measured in 2000–2002. The paper concentrates on factors controlling the winter CO<sub>2</sub> efflux. An extrapolation of soil respiration in a future warmer climate is presented based on soil temperatures.

In **Paper III**, measurements and calculation of an annual CO<sub>2</sub> balance of the Jokioinen peat field growing barley or forage grass are presented. The study covered two years of data. By using the CO<sub>2</sub> exchange and biomass data, the annual net biome production of both crops was calculated.

In **Paper IV** the aim was to determine the annual CO<sub>2</sub> balance of an afforested peat field. In addition, the annual biomass increment was determined based on tree inventory data. Besides the net ecosystem exchange, the paper addresses the most important CO<sub>2</sub> sources and sinks in the ecosystem.



## 1 Introduction

Globally, the soil and vegetation contain nearly three times as much carbon (C) as the atmosphere (Denman et al. 2007). Soil carbon is linked to global climate through microbial processes forming carbon dioxide (CO<sub>2</sub>) and methane (CH<sub>4</sub>), as well as through CO<sub>2</sub> fixed in vegetation by plant photosynthesis. CO<sub>2</sub> and CH<sub>4</sub> are known as greenhouse gases (GHGs); they alter Earth's energy balance by absorbing outgoing longwave radiation. Human activities have greatly increased the concentrations of these gases in the atmosphere during the last few centuries, and especially rapidly during the last few decades, causing climatic warming. The mean atmospheric concentration of CO<sub>2</sub> is currently 379 ppmv and is increasing at an average rate of 1.9 ppmv yr<sup>-1</sup> (Forster et al. 2007).

Mires are peat-forming ecosystems typically characterized by water-saturated conditions and poor oxygen availability, resulting in a decreased decomposition rate of soil organic matter. Accumulation of peat occurs, as the growth of vegetation exceeds the degradation. Despite the fact that mires and peatlands occupy only a few percent of the global land area, the carbon reservoirs in them represent about 20–30% of the Earth's soil C stores (Gorham 1991, Eswaran et al. 1993). Peatlands are found in every climatic zone, but they are particularly abundant in the boreal and sub-arctic regions.

Overall, the net carbon sink constituted by pristine wetlands has diminished globally following their drainage. The exploitation of peat soils has been most vigorous in Europe, where approximately half of the original peatland area has been lost (Joosten and Clarke 2002). On an area basis, one-third of Europe's current peatland resource is located in Finland (Montanarella et al. 2006), where peatlands have been utilized intensively for forestry, agriculture and fuel production (Lappalainen 1996). More than half of the original peatland area (104,000 km<sup>2</sup>) in Finland has been drained, mostly for forestry (57,000 km<sup>2</sup>), but also for agriculture (5000 km<sup>2</sup>) and the peat harvesting industry (530 km<sup>2</sup>) (Vasander 1996). Following the structural changes in agriculture, the area of peatland used for farming has been decreasing: at present there are about 3000 km<sup>2</sup> of croplands that have a soil organic matter concentration exceeding 20%, constituting about 14% of the total area of farmed soils in Finland (Myllys and Sinkkonen 2004). The area of afforested agricultural peat soils has been estimated to be about 1000 km<sup>2</sup> (Wall and Heiskanen 1998).

Owing to the huge amount of C they have accumulated, peatlands play a critical role in the long-term storage of atmospheric carbon and in the global cycling of GHGs. Through their effects on net primary productivity and soil hydrologic conditions, land-use changes and the predicted global warming may fundamentally influence the size of the organic matter pool in the peatlands (Gorham 1991, Martikainen et al. 1995). Hence, these anthropogenic disturbances directly affect the atmospheric concentration of CO<sub>2</sub>, CH<sub>4</sub> and other key GHGs, such as nitrous oxide (N<sub>2</sub>O), and may further accelerate global warming.

Possible options for land-use change to reduce GHG emissions are currently sought. In countries with a high proportion of peatlands, moderate reductions may be achieved by sustaining or increasing the C content in managed peatlands. Feasible methods include

wetland restoration and peatland afforestation (Smith et al. 2007). However, before initiating any actions, it is necessary to assess the GHG implications of different land-use activities, and to gain detailed information of GHG dynamics in managed peatlands.

The objective of this work was to study the atmosphere-biosphere exchange of CO<sub>2</sub> on managed boreal peatlands, and to investigate the environmental and man-made influences controlling it. Measurements of exchange fluxes were conducted on two peatlands with an agricultural history, one of which was still in cultivation, while the other had been afforested 30 years earlier. The emphasis was put on determining the annual CO<sub>2</sub> balance and its components in these ecosystems. More specifically, the aims were

- to test the suitability of a response model for separating the soil respiration from the total ecosystem respiration
- to compare annual CO<sub>2</sub> emissions from soil respiration on mineral and peat soils
- to study the dynamics of wintertime CO<sub>2</sub> exchange on an agricultural peat soil
- to determine the CO<sub>2</sub> fluxes and annual balances on an agricultural peat soil with two different crops
- to examine the effect of afforestation on the annual CO<sub>2</sub> balance of an organic-soil cropland

In this thesis summary, the biological and physical components controlling the atmosphere-biosphere CO<sub>2</sub> cycle (Chapter 2), and the carbon dynamics in northern peatlands (Chapter 3) are first introduced. In Chapter 4, the measurement sites and methodology are presented. Chapter 5 summarizes the main results of the study as presented in the research papers included in the thesis. In addition to these results, some new data on soil CO<sub>2</sub> efflux are presented that help in understanding the role of the soil respiration in the carbon balances of the studied sites.

## 2 Controls on net ecosystem CO<sub>2</sub> exchange

### 2.1 Biological processes

The ecosystem carbon balance consists of several processes having opposite direction (Fig. 1). CO<sub>2</sub> is taken up in the photosynthesis of plants (gross primary production, GPP) which is the main process transporting carbon from the atmosphere to the biosphere. At the same time, autotrophic plant respiration ( $R_d$ ) produces and releases CO<sub>2</sub> into the atmosphere. The net flux resulting from these two processes is called the net primary production (NPP):

$$\text{NPP} = \text{GPP} + R_d \quad (1)$$

NPP is equal to the plant growth and biomass increment. In ecological concepts GPP typically has values above zero, denoting uptake of CO<sub>2</sub> by plants from the atmosphere. Here, however, it takes negative values, denoting removal of CO<sub>2</sub> from the atmosphere.

A CO<sub>2</sub>-producing community of microbes and macrofauna present in the soil is responsible for the decomposition of dead organic matter. The CO<sub>2</sub> flux arising from these sources is called the heterotrophic soil respiration ( $R_s$ ). Summing this with NPP yields the net ecosystem production (NEP), which is equal to the net ecosystem CO<sub>2</sub> exchange (NEE):

$$NEE = NEP = NPP + R_s \quad (2)$$

The NEE taken over a whole ecosystem, and integrated over time, defines the loss or gain of CO<sub>2</sub>, thus determining whether the ecosystem is a source or a sink of CO<sub>2</sub>. Both NEE and NEP are typically used to cover wide temporal ranges; however, NEE often refers to a short-term (30-minute) ecosystem-atmosphere flux. In this work, an atmospheric view of fluxes has been adopted: negative NEE values denote removal of CO<sub>2</sub> from the atmosphere into the ecosystem, whereas positive NEE values indicate the reverse.

Accounting for longer-term disturbances, i.e., non-respiratory processes leading to loss of living and dead organic material, such as harvest, erosion, leaching, forest cutting or fire, results in the net carbon balance. This is defined as the net biome production (NBP):

$$NBP = NEP + \sum_{i=1} D_i \quad (3)$$

where  $D_i$  ( $i = 1, 2, \dots$ ) denotes a long-term disturbance.

### 2.1.1 Photosynthesis (GPP)

In photosynthesis, atmospheric CO<sub>2</sub> is converted into more complex organic compounds by the light energy from the sun. While different plant species vary greatly in their photosynthetic efficiency, the external factors governing GPP are universal: temperature, irradiance, CO<sub>2</sub> concentration and soil nutrient and water conditions (Weier et al. 1982).

The temperature response of GPP typically has an optimum at 15–25 °C for many boreal species. In forests, the temperature sensitivity of GPP is high in the range 0–15 °C, but decreases above that (Grace 2005). At low temperatures, at or below zero, photosynthesis is suppressed due to the slowing down of enzymatic reactions. The relationship between photosynthesis and radiation is often described by a rectangular hyperbola, rising steeply from darkness until a saturation level is reached, where a further increase in photon flux does not increase photosynthesis (Ruimy et al. 1995).

Nitrogen is essential in the synthesis of chlorophyll, the molecule involved in photosynthesis. At the ecosystem level, GPP is correlated with the amount of green leaf area (Aurela 2005, Hyvönen et al. 2006) and leaf nitrogen content (Tieszen and Detling 1983, Schulze et al. 1994).

Water is needed in the process of photosynthesis. However, 99% of the water taken up by the plant from the soil flows through the plant from roots to leaves, and is ultimately transpired into the atmosphere through the stomata (Weier et al. 1982). The stomata link the

water and carbon cycle by controlling the gas exchange and thus the diffusion of CO<sub>2</sub> into the leaf. In a situation of water deficiency, the stomata close to prevent excess water loss, thus markedly affecting the CO<sub>2</sub> uptake (Waring and Running 2007).

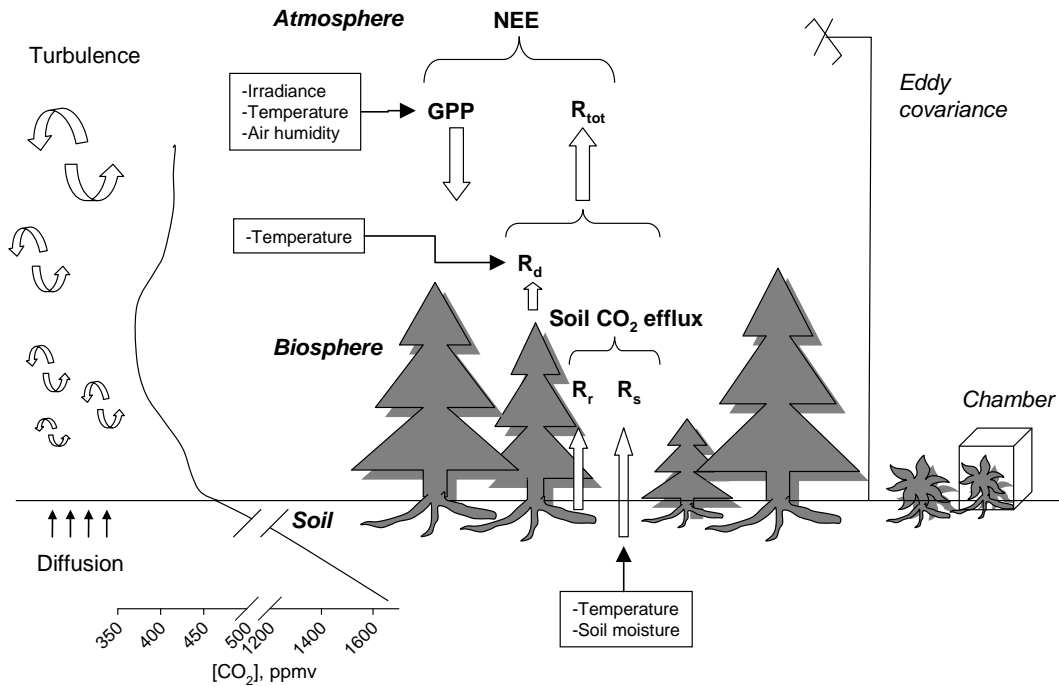


Figure 1. Schematic view of the transport mechanisms of gases (on the left-hand side of the figure), CO<sub>2</sub> flux components (middle), the most important environmental factors controlling them (in boxes), and the flux measurement methods used in this study (right). Also shown is a typical CO<sub>2</sub> concentration profile in the soil-air continuum during daytime in well-mixed atmospheric conditions with photosynthetic uptake by the forest canopy. NEE = net ecosystem exchange, GPP = gross primary production,  $R_{tot}$  = total ecosystem respiration,  $R_s$  = soil respiration,  $R_d$  = plant respiration,  $R_r$  = rhizosphere respiration.

### 2.1.2 Total ecosystem respiration ( $R_{tot}$ )

The total ecosystem respiration ( $R_{tot}$ ), a sum of  $R_d$  and  $R_s$ , integrates the heterotrophic and autotrophic respiration processes of the ecosystem. Beside the functional classification, it is sometimes useful to divide the respiratory CO<sub>2</sub> fluxes according to their spatial origin. Then,  $R_{tot}$  is comprised of above- and below-ground components, the latter, including  $R_s$  together with the root and rhizosphere respiration ( $R_r$ ), equals the observed soil CO<sub>2</sub> efflux (Fig. 1). Functionally, however,  $R_r$  is part of  $R_d$ . In practice, partition between autotrophic and heterotrophic soil respiration is complex and technically demanding (Trumbore 2006). In the following, a short description of each respiration component and its controlling factors is given with mixed functional and spatial classifications.

### **2.1.2.1 Autotrophic respiration ( $R_d$ )**

25–75% of the CO<sub>2</sub> fixed by photosynthesis is consumed in autotrophic respiration (Ryle 1984), which is essential to provide metabolic energy and carbon skeletons for plant growth and the maintenance of plant tissue. Conceptually,  $R_d$  is often separated into maintenance and growth respiration (McCree 1974). Experimentally,  $R_d$  has been shown to relate to canopy temperature and the growth stage, crop canopy respiration being highest at or just after the heading of the crop (Mogensen 1977, Baldocchi 1994). In productivity and carbon cycle models,  $R_d$  is typically exponentially increased by temperature (Adiku et al. 2006).

### **2.1.2.2 Soil heterotrophic respiration ( $R_s$ )**

Soil organic matter is a mixture of decomposing plant residues, the microorganisms themselves, and resistant humic substances. The decomposition of organic matter, i.e., soil heterotrophic respiration (hereafter referred to as soil respiration) is carried out by fungi, bacteria and soil animals, whereas the reactions are mediated by enzymes. Microorganisms use carbon compounds for building new cells and as an energy source (Paul and Clark 1989).

$R_s$  is controlled by a wide range of different environmental drivers. The most important factors are soil temperature (Kim and Verma 1992, Davidson and Janssens 2006), and soil moisture (Moore and Dalva 1993, Silvola et al. 1996a). Typically  $R_s$  increases exponentially with temperature (Lloyd and Taylor 1994). In peat soils, the soil moisture is closely linked to water table depth, which determines the thickness of the oxic layer. Water table depth, on the other hand, is coupled with temperature, since high temperatures promote evapotranspiration, which lowers the water table.

In addition to climatic factors, primary production (Raich and Schlesinger 1992) and the quality and quantity of the organic matter (La Scala et al. 2000) have an influence on the soil respiration. In forestry-drained peat soils, the nutrient concentration has been shown to have a positive correlation with peat decomposition rates (Silvola et al. 1996a, Minkinen et al. 2007). The carbon-nitrogen ratio of soil approximately describes the quality of the soil organic matter. In nutrient-poor pristine peatlands, C/N ratio is typically high, but decreases in drained and managed peat soils through mineral soil addition, ploughing and fertilization, as the mineralization rates are increased.

Although strongly influenced by temperature, soil CO<sub>2</sub> production does not cease during the wintertime, but is potentially an essential part of the annual CO<sub>2</sub> emission in snow-covered ecosystems (Koizumi et al. 1996, Alm et al. 1999, Fahnestock et al. 1999, Aurela et al. 2002, Schindlbacher et al. 2007). In boreal peat soils, the thermal insulation of the soil by the snow cover and the stabilizing effect of water due to its high heat capacity typically maintain near- or above zero temperatures in the soil, and enable microbial processes to continue throughout the year. In addition, microbial production and emission of CO<sub>2</sub> has been shown to occur also in frozen soil (Coxson and Parkinson 1987, Clein and Schimel 1995), even in temperatures as low as -17 °C (Panikov and Dedysh 2000).

### **2.1.2.3 Rhizospheric respiration ( $R_r$ )**

In typical soil CO<sub>2</sub> efflux studies, the measured flux represents both  $R_s$  and root and rhizospheric respiration ( $R_r$ ).  $R_r$  includes the respiration of roots and their associated mycorrhizal fungi and other micro-organisms in the rhizosphere directly dependent on root exudates. It is a significant part of the soil CO<sub>2</sub> efflux, as even more than half of the carbon fixed in photosynthesis may be released by  $R_r$  (Kuzyakov and Cheng 2001, Högberg et al. 2001, Silvola et al. 1996b).

In a wide range of different ecosystems, root and rhizosphere respiration have been shown to be tightly coupled with plant photosynthesis, with similar seasonal cycles (Silvola et al. 1996b, Larsen et al. 2006, Sampson et al. 2007). Comprehensive separation of root respiration from the other rhizospheric sources (soil, root exudates, etc.) has proved difficult, although many different approaches, like root exclusion by trenching or girdling, or non-destructive methods like isotope techniques and modelling have been used (Subke et al. 2006). Despite the considerable attention recently given to  $R_r$ , in this work it has not been treated separately but is defined as being part of  $R_s$ .

## **2.2 Physical transport mechanisms**

NEE measured above the ecosystem is not only a function of the biological processes, but also of the physical phenomena affecting the gas transport in the soil, the atmosphere and at the vegetation-atmosphere interface. The most important transport mechanisms in the atmosphere-biosphere exchange are atmospheric turbulence and molecular diffusion, the latter of which arises from the Brownian motion of molecules. These mechanisms are presented shortly below, together with some examples linking them to the experimental methods employed in this study.

### **2.2.1 Turbulence**

The troposphere, which is the lower part of the atmosphere, may be divided into the free atmosphere and the atmospheric boundary layer (ABL), the latter being directly affected by the underlying surface. The height of the ABL typically ranges from 100 m to 2 km (Stull 1988). The lowest 10% of the ABL is called the surface layer (SL). In the bulk of the ABL, the vertical transport of gases is governed by turbulent mixing. Turbulence consists of many different sizes of irregular swirls of motion, called eddies, that transport material upwards and downwards. In the SL, the vertical variation in the turbulent fluxes is small, meaning that a flux measured within the SL approximately equals that at the surface (Stull 1988). Turbulence is generated by friction (mechanical turbulence) and surface heating (thermal turbulence).

Turbulence plays an important role in gas flux measurement techniques. The eddy covariance (EC) method is based on the detection of turbulent fluctuations and depends, for example, on an assumption of well-developed turbulence. During low wind conditions, a correlation between the intensity of turbulence and the CO<sub>2</sub> efflux is commonly observed

(Goulden et al. 1996, Aurela 2005). This indicates a potential underestimation of the CO<sub>2</sub> efflux, particularly during calm nights. Biases caused by this phenomenon are typically corrected by replacing the flux values measured in weak turbulence with those derived from measurements in well-mixed conditions (Baldocchi 2003).

In closed chamber measurements, turbulence is generated artificially by fans located inside the chamber. Fans are needed to ensure proper mixing of the gases in the chamber. With no mixing, a vertical concentration gradient is developed inside the chamber, slowing down the diffusion from the soil. Furthermore, adequate mixing of gases is a prerequisite for taking representative samples from the chamber headspace. On the other hand, too high a mixing rate may induce an artificially high burst of CO<sub>2</sub> from the soil, leading to an overestimated efflux rate (Le Dantec et al. 1999, Pumpanen et al. 2004).

### **2.2.2 Molecular diffusion**

Production of CO<sub>2</sub> in the soil generates a vertical concentration gradient (Fig. 1). In the soil and within the first few centimetres above it, molecular diffusion is a significant mechanism of gas transport. According to Fick's law, the flux rate of CO<sub>2</sub> is equal to the concentration gradient multiplied by the diffusion coefficient. As the diffusion rate of gases is about 10,000 times higher in air as compared to water, the effectiveness of the diffusive transport depends largely on the soil porosity and the water-filled pore space.

The large difference in the gaseous diffusion rate between air and water has an important implication in peatlands. Because of the slow diffusion, the water-saturated peat is largely depleted in oxygen, due to which the decomposition of organic material in the anoxic conditions is low. Consequently, accumulation of peat takes place. When the water table drops following wetland drainage, enhanced gaseous diffusion leads to the rapid decomposition of peat, releasing large quantities of CO<sub>2</sub> (Silvola et al. 1996a)

In practice, there is also non-diffusive transport of CO<sub>2</sub> from the soil into the atmosphere. This mass movement is induced by the wind and atmospheric pressure changes (Massman et al. 1997). At worst, this may lead to a poor correlation between the CO<sub>2</sub> efflux measured with chambers and the CO<sub>2</sub> production calculated using the soil concentration profile and diffusion coefficients (Risk et al. 2002). The significance of the advection-induced transport is likely to increase in dry and porous soils (Risk et al. 2002, Hirsch et al. 2004).

The freezing of soil water may produce a physical barrier for the gas exchange, leading to reduced diffusion rates and accumulation of greenhouse gases in the soil with a subsequent release of trapped gases at soil thawing (Friborg et al. 1997). It is not only the diffusion of CO<sub>2</sub> out of the soil, but also the diffusion of molecular oxygen from the atmosphere into the soil which may be hindered. This may result in suppression of the soil respiration. The observed fluxes above the soil are then decreased due to both biological and physical phenomena.

### **3 Carbon dynamics in natural and agricultural peat soils**

#### **3.1 Natural peat soils**

Boreal peatlands began to develop after the end of the last glacial period, about 10,000 years ago. Since the decomposition of organic matter in pristine mires is typically slower than the input of carbon by the wetland vegetation, these ecosystems have been accumulating biomass carbon. A long-term carbon accumulation rate of approximately  $19 \text{ g C m}^{-2} \text{ yr}^{-1}$  has been measured in Finnish pristine mires (Turunen et al. 2002). At present, direct year-round  $\text{CO}_2$  and  $\text{CH}_4$  exchange measurements conducted with the EC method on several temperate, boreal and subarctic wetlands show an uptake of carbon, annual  $\text{CO}_2$  balances ranging from  $-10$  to  $-75 \text{ g C m}^{-2}$  (Lafleur et al. 2003, Aurela et al. 2004, Aurela et al. 2007a, Lund et al. 2007), with methane emissions ranging from  $4$  to  $10 \text{ g C m}^{-2}$  (Hargreaves et al. 2001, Rinne et al. 2007). However, a literature review of gas balances in boreal mires obtained by measurements and modelling indicated much more variable flux rates, with annual  $\text{CO}_2$  balances varying from  $-98$  to  $101 \text{ g C m}^{-2}$  (Saarnio et al. 2007). The annual losses of  $\text{CO}_2$  were generally observed during dry years, highlighting the strong control hydrology typically exerts on gas balances.

To estimate the carbon balance of a peatland, besides net ecosystem exchange, one needs to know e.g. the leaching and the airborne erosion of carbon. In peatlands, most of the carbon moving with the water occurs as dissolved organic carbon (DOC). Leaching of DOC takes place both laterally, in and out of the system along the surface or in ground water, as well as vertically, leaching down to the mineral soil underlying the peat (Domisch et al. 1998, Moore 2003).

In general, the amount of exported DOC is correlated with rainfall, and thus the amount of runoff water (Scott et al. 1998), and the net primary productivity. The draining of a peatland does not necessarily change carbon leaching dramatically: similar rates of DOC export, about  $10 \text{ g C m}^{-2} \text{ yr}^{-1}$  were reported for pristine and forestry-drained peatlands in central Finland (Sallantausta 1994). Leaching of DOC may actually be greater from improperly drained areas, for example when the ditches get choked, owing to longer water retention time in the peat (Scheithauer and Grunewald 2004). However, a leaching rate of  $35 \text{ g C m}^{-2} \text{ yr}^{-1}$  has been reported for peatlands in North America with a higher flow-through (Dosskey and Bertsch 1994). Airborne erosion of carbon may be significant in cultivated peat fields, mainly during the periods when there is no vegetation on the field and the soil is being ploughed, but in peatland forests it may be assumed negligible. In an upland moorland in the UK, the amount of peat eroded by the combined effects of wind and rain was about  $50 \text{ g m}^{-2} \text{ yr}^{-1}$  (Warburton 2003).

#### **3.2 Peat soils with an agricultural history**

By disturbing the natural hydrological conditions and nutrient and gas fluxes, the management of peatlands by draining alters their carbon cycling. Characterized by low bulk



density and high porosity, peat undergoes subsidence (Armentano and Menges 1986, Minkkinen and Laine 1998a, Schipper and McLeod 2002). This is first caused by the physical collapse of the peat matrix and later on by the microbial decomposition of the peat (Päivänen and Paavilainen 1996). In the long run, the peat deposit gets thinner and starts to mix in with the underlying mineral soil, changing gradually to mull. Despite the compaction, the thermal conductivity in the peat is significantly lower than in mineral soils (DeVries 1975), leading to a smaller diurnal amplitude in the soil temperature as compared to non-peaty soils. As a practical consequence, peat soils freeze later in the winter and the frozen layer may remain thinner as compared to mineral soils (Mylly 1996).

Following the water-table drop, the gas exchange in peatlands is affected significantly (Freeman et al. 1992). The exposure of the peat layer to oxygen intensively stimulates the mineralization of the organic material. The fertility and trafficability of the drained peat soils have often been improved by adding a mineral soil admixture into the soil, which further enhances the aeration of the peat. As a consequence, the high CH<sub>4</sub> emissions typical of pristine peatlands cease; drained peatlands may even turn into a small sink of CH<sub>4</sub> (Maljanen et al. 2003). By contrast, the CO<sub>2</sub> and N<sub>2</sub>O emissions from such peat soil are generally greatly enhanced (Martikainen et al. 1993, Nykänen et al. 1995, Regina et al. 1996, 2004).

Even in mineral soils brought into agricultural use, the carbon content generally decreases. This is a result of two separate processes: first, the soil respiration is increased. The decomposition rates in farmed soils are generally higher than in forest soils due to the higher soil temperature, the more easily decomposable types of litter, the greater physical accessibility of soil organic matter to decomposing organisms due to tillage, the better soil aeration and more optimal moisture conditions there. Second, the litter input is decreased, as most of the biomass is harvested (Schlesinger and Andrews 2000).

As compared to mineral soils, the carbon losses in drained peat soils are much greater, since the decomposition rate of carbon-rich peat is accelerated owing to the oxic conditions, ploughing and fertilization. Substantial carbon losses of up to 750 g C m<sup>-2</sup> yr<sup>-1</sup> have been reported in farmed peat soils (Nykänen et al. 1995, Maljanen et al. 2001). CH<sub>4</sub> fluxes in cultivated peat soils are typically low, annual balances varying from -0.12 to 4.0 g C m<sup>-2</sup> yr<sup>-1</sup> in Finland (Regina et al. 2007). N<sub>2</sub>O emission rates from cultivated peat soils are considerable, with annual emissions from different crops ranging from 0.73 to 2.5 g N m<sup>-2</sup> yr<sup>-1</sup> (Regina et al. 2004).

Afforestation of a former agricultural mineral soil greatly increases the carbon store in the above-ground vegetation, whereas the amount of C sequestered in the soil is small (Paul et al. 2002). According to the review by Post and Kwon (2000), the increase in soil C following afforestation averaged 32 g C m<sup>-2</sup> yr<sup>-1</sup> across different climatic zones, the variation being large (from -90 to 310 g C m<sup>-2</sup> yr<sup>-1</sup>). Despite the small C accumulation in the long term, the C content of the upper horizons may even diminish during the first decades after the afforestation (Vesterdal et al. 2002, Huang et al. 2007). In general, the input of new organic matter into the soil is greater in forest ecosystems with much longer rotation periods as compared to agricultural soils (Jandl et al. 2007).

Recently, soil CO<sub>2</sub> fluxes in afforested and otherwise managed peat soils have been studied intensively (Byrne and Farrell 2005, Mäkiranta et al. 2007, Minkkinen et al. 2007), but these do not include the contribution of the forest canopy. Cannell et al. (1993) have estimated that planting conifer forests on drained peatlands leads to CO<sub>2</sub> accumulation, if the underlying peat layer is not thicker than 15–40 cm. Hargreaves et al. (2003) estimated, based on combined measurements and modelling, that an afforested peatland (with no agricultural history) turns into an annual sink of CO<sub>2</sub> 4–8 years after the afforestation. Evidence obtained by peat coring shows that in nutrient-poor forested peatlands the carbon store in peat has increased several decades after drainage, whereas more fertile sites have experienced carbon loss (Minkkinen and Laine 1998b). However, direct flux measurements on afforested organic-soil croplands have not been published earlier.

## 4 Materials and methods

### 4.1 Field sites

The eddy covariance CO<sub>2</sub> exchange measurements on cultivated peat soil presented in **Papers II and III** were conducted from October 2000 to May 2003 at Jokioinen (60°54'N, 23°31'E, 104 m above sea level) (Fig. 2). In May 2001, the field was planted with spring barley (*Hordeum vulgare*) and undersown grass (a mixture of *Phleum pratense* and *Festuca pratensis*). The barley was harvested in September 2001. In 2002, forage grass was grown on the site, with two harvests taking place in the latter half of June and in mid-August. The depth of the peat layer (Terric histosol) at the site was about 0.5 m. This layer contained carbon on average 56 kg C m<sup>-2</sup> within a distance of 120 m from the measurement masts. The maximum measured projected leaf area index (LAI) for barley and grass were 6 and 5.5 m<sup>2</sup> m<sup>-2</sup>, respectively.

Measurements of CO<sub>2</sub> fluxes by chambers and soil CO<sub>2</sub> concentration (**Papers I and II**) were made in 2000–2002 on the adjacent experimental field about 30 m to the northwest of the EC masts. This plot consisted of nine subplots (10 m × 10 m) growing barley (*H. vulgare*), forage ley (a mixture of *P. pratense* and *F. pratensis*) and potato (*Solanum tuberosum*), three replicate subplots of each. The average carbon content of the 0.5 m peat layer was 60 kg C m<sup>-2</sup>. Chamber measurements were also made on clay (Vertic cambisol) and loamy sand (Eutric cambisol) soils (see details in Table 1 in **Paper I**). The experimental fields with peat, clay or sandy soils were located no more than 15 km from each other.

The CO<sub>2</sub> exchange on an afforested peat field described in **Paper IV** was measured at Alkkia (municipality of Karvia) (62°11'N, 22°47'E, 159 m above sea level) (Fig. 2). The site, originally drained in 1936–1938, was in agricultural use for 30 years, after which, in 1971, it was planted with Scots pine (*Pinus sylvestris*) with row spacings of 3, 5 and 8 m. The mean diameter (± standard deviation) of the trees at the breast height was 14.8 ± 5.9 cm and height 12.1 ± 2.2 m. The total stand density was 977 ± 393 trees ha<sup>-1</sup> and stand volume 126 ± 72 m<sup>3</sup> ha<sup>-1</sup>. Because of the varying row spacing, the LAI of the forest canopy varied

from 0.7 to 2 m<sup>2</sup> m<sup>-2</sup>. On the forest floor the vegetation was dense, the LAI averaging 2 m<sup>2</sup> m<sup>-2</sup>. The mean peat depth, measured at 12 points within a 200-m radius of the EC mast, was 1.56 ± 0.13 m.

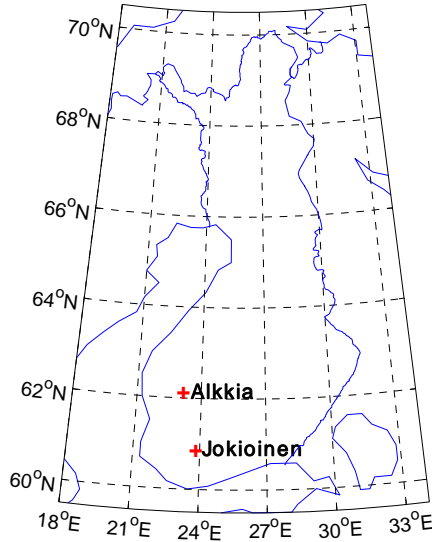


Figure 2. Locations of the measurement sites of Jokioinen (*Papers I–III*) and Alkkia (*Paper IV*) in Finland.

Table 1. Long-term climatic parameters (means of the 1971–2000 period) and soil characteristics at the Jokioinen and Alkkia sites.

	<b>Cultivated peat field (Jokioinen)</b>	<b>Afforested peat field (Alkkia)</b>
<i>Climate</i>		
Annual mean temperature (°C)	4.3	3.1
Annual precipitation (mm)	607	654
Snow depth in mid-March (cm)	31	40
<i>Soil characteristics</i>		
Mean peat depth (m)	0.5	1.6
Water table depth (m)	> 0.8	0.4–0.6
C content (%) <sup>a</sup>	24	16
N content (%) <sup>a</sup>	1.1	0.6
C/N ratio <sup>a</sup>	21	26
pH <sup>a</sup>	5.8	4.7
Bulk density (g cm <sup>-3</sup> ) <sup>a</sup>	0.48	0.62

<sup>a</sup>at Jokioinen, in the top 25 cm of the soil; at Alkkia, in the top 20 cm of the soil, litter removed

The annual mean temperature was higher and the precipitation was lower on the cultivated field as compared to the afforested peat field (Table 1). The soil pH in the first 20–25 cm layer was also higher in the former but the C/N ratio was lower, reflecting the higher fertility. The thickness of the peat at Alkkia was thrice that at Jokioinen, whereas the ground water level was closer to the soil surface at Alkkia.

## 4.2 Flux measurement techniques

In three papers (**Papers II–IV**) out of the four in this thesis, the atmosphere-biosphere CO<sub>2</sub> exchange was measured with the micrometeorological eddy covariance method. The total ecosystem respiration was analyzed with the chamber method in **Papers I and II**. Some additional chamber measurements from bare soils are also included in this thesis summary.

### 4.2.1 Eddy covariance method

The EC method is based on observing the motions and concentrations of upward and downward moving air to determine the net flux of the constituent under examination, in this case CO<sub>2</sub> (Baldocchi 2003). In practice, the atmospheric concentration of CO<sub>2</sub> ( $c$ ) is measured above the studied ecosystem at a high frequency (typically 10 Hz) simultaneously with the vertical wind velocity ( $w$ ). The vertical mass flux density ( $F_c$ ) of CO<sub>2</sub> is then obtained as the covariance between instantaneous fluctuations in  $w$  and  $c$ . Although the raw data is collected at a high frequency, the flux is calculated for a longer time period, generally 30 minutes. This is done to cover the wide frequency range contributing to the turbulent transport, extending from approximately 0.001 to 10 Hz.

Under ideal conditions it may be assumed that  $F_c$  equals the biological sink/source strength of the surface under examination. In such conditions, the storage of CO<sub>2</sub> below the measurement height should be zero. In addition, the surface under examination is assumed to be horizontally homogenous and flat. Then, no vertical or horizontal advection of CO<sub>2</sub> should occur to disturb the mass balance assumed for the EC method (Baldocchi 2003).

Including all sources and sinks below the measurement height, the flux of CO<sub>2</sub> obtained at an EC measurement is equal to the short-term net ecosystem exchange (NEE). Horizontally, the flux represents the average exchange from an area upwind of the measurement point. This area is called the flux source area, or footprint, and the location and extent of this area depends on the atmospheric stability, surface roughness, wind speed and the measurement height (Schmid 2002).

In practice, the requirements for ideal conditions are never fully achieved; additionally, several post-processing operations and corrections are needed when calculating the EC flux from the high-frequency data collected. These include data despiking, detrending, determining the lag-time between the  $w$  (sonic anemometer) and  $c$  (CO<sub>2</sub> analyzer) signals, coordinate rotation and correction of density fluctuations (Webb-Pearman-Leuning correction) (Webb et al. 1980), correction for the attenuation of low and high frequency

contributions to the flux, and screening for non-physical values (Baldocchi 2003, Aurela 2005, **Paper III**).

During conditions of suppressed turbulence, typically on calm summer nights, the net flux density of CO<sub>2</sub> moving into and out of the plant-soil system can be larger than the measured flux because of the accumulation of CO<sub>2</sub> in the air layer below the measurement height (Aubinet et al. 2000). The opposite phenomena takes place at sunrise when the stable nocturnal boundary layer breaks up and the stored CO<sub>2</sub> is released and mixed with the upper air layers. The magnitude of the storage flux can be estimated by measuring the vertical concentration profile below the EC measurement (**Paper IV**).

In micrometeorology, an important quantity describing the intensity of mechanically-generated turbulence is the friction velocity ( $u^*$ ). In stable conditions, the measured flux  $F_c$  tends to increase along with  $u^*$ , even after accounting for the storage term (Massman and Lee 2002, **Papers III–IV**), causing underestimation of the fluxes, especially at night-time. This error may be avoided by replacing the flux values measured during periods of insufficient turbulent mixing with values obtained from the temperature dependence of the data measured during sufficiently strong turbulence. The limit for sufficient turbulence is typically determined empirically and individually for each data set; no universal limits exist (Gu et al. 2005).

In this thesis, the instrumentation for the EC measurements included an SATI-3SX sonic anemo/thermometer (Applied Technologies, Inc.) (**Papers II–IV**). The fast-response infrared (IR) CO<sub>2</sub>/H<sub>2</sub>O analyzers used were an LI-6262 (Li-Cor, Inc.) at Jokioinen (**Papers II and III**) and an LI-7000 at Alkkia (**Paper IV**). The measurement heights were 3 and 18 m at Jokioinen and Alkkia, respectively. Data sampling was conducted at both sites at a rate of 10 Hz.

#### **4.2.2 Chamber method**

In this work, both static and dynamic closed chamber methods were used. This division is based on the presence or absence of the through-flow air circulation (Witkamp and Frank 1969). In both closed chamber methods, the concentration increase inside a sealed chamber is monitored, typical closure times varying from a few minutes to a few hours. In closed static chambers, sampling may be done using an absorbing media like soda lime (Janssens and Ceulemans 1998), or by taking separate samples from the chamber headspace at even intervals and analyzing the samples elsewhere (Regina et al. 2004).

In the case of the closed dynamic chamber technique, on the other hand, the sample air is sucked continuously from the chamber; either it is then returned to the chamber headspace after being analyzed, or the sample air is replaced with ambient air. There exist numerous chamber types with characteristic design and technical details, all having their advantages and disadvantages (Fang and Moncrieff 1996, 1998, Rayment and Jarvis 1997, Rochette et al. 1997, Le Dantec et al. 1999, Janssens et al. 2000, Longdoz et al. 2000, Pumpanen et al. 2001). A thorough comparison of twenty different chambers representing different designs and measurement principles against a known CO<sub>2</sub> flux, however, suggested that none of the

different measurement principles proved to be superior in all respects (Pumpanen et al. 2004).

In this thesis, altogether 4 chamber assemblages were used in 5 measurement campaigns (Table 2). All chamber CO<sub>2</sub> flux measurements were performed on collars, which were inserted into the soil well before the measurements. In Experiment A, additional steel collars (20 and 40 cm) were used to fit the vegetation inside the chamber. With the large cubic chambers (Exp. A and D), water was used to make the chamber air-tight. All chambers were equipped with a fan. Measurements were conducted on vegetation-covered soil (Exp. A), on bare ploughed soil (Exp. C) and on soil from which the above-ground vegetation was removed by clipping (Exp. B and E). With all chamber types, the flux rate was calculated from the linear increase in the chamber headspace CO<sub>2</sub> concentration. The soil temperature was typically measured at a depth of 4–5 cm.

To estimate the annual carbon loss from bare soil, hitherto unpublished CO<sub>2</sub> efflux data are introduced in this thesis summary. Measurements were conducted in cultivated peat, clay and sandy soils (Exp. C) and in afforested peat field (Exp. E). At the cultivated sites, measurements were made on bare fallow plots which were kept free of vegetation. Three collars (diameter 10 cm, height 5 cm) were inserted into the soil on two replicate subplots. Measurements of  $R_s$  were made approximately twice a week during the growing season (Exp. C) and weekly to biweekly in winter (Exp. D). The wintertime (October–March) emission was calculated from the monthly means of the measured CO<sub>2</sub> fluxes. For the growing season (April–September), a temperature response function was fitted to the measured data, and the cumulative emission was estimated using the obtained parameter values and continuous soil temperature data.

For the afforested peat field, data published by Mäkiranta et al. (2007) were used for comparison (Exp. E). In this experiment, the above-ground vegetation and the litter were removed and the formation of new litter was prevented. In addition, the roots were cut at a depth of 30 cm with a metal cylinder. It was assumed that, after this manipulation, the observed soil CO<sub>2</sub> efflux only represents the soil respiration, i.e., the decomposition of the peat. Data from 2001–2002 and 2003–2004 were employed for cultivated and afforested peat fields, respectively.

### **4.3 Parameterization of the fluxes**

In this work, parameterization of the fluxes was performed in order to (1) compare the soil respiration rates in different soil and crop types by separating the autotrophic and heterotrophic components of the total ecosystem respiration by modelling, (2) calculate the annual CO<sub>2</sub> emission from measurements of  $R_s$  on different soil types and (3) fill in the gaps in the EC flux data arising from e.g. instrument malfunctions, enabling the calculation of the seasonal and annual balances.

Table 2. The chamber types, chamber measurement campaigns and sites included in this work with references to presented results.

Experiment	Site <sup>b</sup>	Time	Respiration component	Soil types	Plant cover	Chamber method <sup>c</sup> and closure time	Chamber type and size <sup>d</sup>	Frequency of measurements	Analyzer type and name	Manipulation of plots	Results shown in and references
A	Jok	May–Sept 2000	$R_{tot}$	Peat Sandy Clay	Barley Grass Potato	CD 1–2 min	Cubic aluminium chamber $A = 60 \text{ cm} \times 60 \text{ cm}$ $H = 20 \text{ cm}$	Weekly	IR LCA-2		Chapter 5.1.1 <b>Paper I</b>
B	Jok	Spring/ Autumn 2000	$R_s$	Peat Sandy Clay	Barley Grass Bare soil	CD 1–2 min	Cylindrical SRC-1 chamber PP Systems $\varnothing = 10 \text{ cm}$ $H = 5 \text{ cm}$	Spring: 2–3 times Autumn: 1–2 times per week	IR EGM-2	Above-ground vegetation clipped regularly	Chapter 5.1.1 <b>Paper I</b>
C	Jok	Growing season 2002	$R_s$	Peat Sandy Clay	Bare soil	CD 1–2 min	Cylindrical SRC-1 chamber PP Systems $\varnothing = 10 \text{ cm}$ $H = 5 \text{ cm}$	2–3 times per week	IR EGM-2	Bare ploughed soil	Chapter 5.1.2, Table 3
D	Jok	Winters 2000–2001 and 2001–2002	$R_s$	Peat	Barley Grass Bare soil	CS 30 min	Cubic aluminium chamber $A = 60 \text{ cm} \times 60 \text{ cm}$ $H = 20 \text{ cm}$	Weekly to monthly	FID HP 6890 Series		Chapter 5.1.2, Table 3 <b>Paper II</b> Regina et al. 2004
E <sup>a</sup>	Alk	2003–2004	$R_s$ (peat only)	Peat	Peat only	CD 80 s	Cylindrical aluminium chamber $\varnothing = 15.7 \text{ cm}$ $H = 14.9 \text{ cm}$	Weekly to monthly	IR EGM-4	Vegetation and litter removed and roots cut	Chapter 5.4.1 Chapter 5.1.2, Table 3 <b>Paper IV</b>

<sup>a</sup>Mäkiranta et al. 2007

<sup>b</sup>Jok = Jokioinen, Alk = Alkkia

<sup>c</sup>CD = closed dynamic, CS = closed static

<sup>d</sup> $A$  = basal area,  $H$  = height,  $\varnothing$  = diameter of the chamber

The division of  $R_{tot}$  into  $R_s$  and  $R_d$  was done in **Papers I and III**, whereas in **Papers II and IV** no partition was done, and  $R_{tot}$  was used in the modelling. In all of the papers, and when calculating the cumulative soil CO<sub>2</sub> loss from the new data, the equation for soil respiration (for  $R_{tot}$  in **Papers II and IV**) was adopted from Lloyd and Taylor (1994):

$$R_s = R_{s0} \times \exp \left[ E_0 \times \left( \frac{1}{T_l} - \frac{1}{T_s - T_{0,s}} \right) \right] \quad (4)$$

where  $R_{s0}$  is a parameter describing the soil respiration (in mg CO<sub>2</sub> m<sup>-2</sup> s<sup>-1</sup>) at the reference soil temperature of 273.15 or 283.15 K, depending on the site and season, and  $T_l$  is the respective reference temperature (46.02 or 56.02 K).  $T_s$  denotes the soil temperature at a selected depth (in K), and  $T_{0,s} = 227.15$  K.  $E_0$  is a parameter describing the temperature response of the soil respiration. For  $E_0$ , both a constant value (**Papers I–III**) and a value derived from the data (**Papers II and IV**) were used, depending on the application. The constant value equalled 308.56 K, as in Lloyd and Taylor (1994). In **Paper I**, a semi-empirical coefficient called effective soil activity was used to account for the activity of recently-fixed carbon in  $R_s$ . In **Papers I and III**, the autotrophic respiration was defined separately with a temperature-response equation resembling Eq. (4).

The parameterization of the gross photosynthesis was based on a commonly-used nonlinear radiation response function (Ruimy et al. 1995):

$$NEE = GPP + R_{tot} = \left( \frac{\alpha \times \text{PPFD} \times GPP_{\max}}{\alpha \times \text{PPFD} + GPP_{\max}} \right) + R_{tot} \quad (5)$$

where  $\alpha$  is the apparent quantum yield (in mg  $\mu\text{mol}^{-1}$ ),  $GPP_{\max}$  (mg m<sup>-2</sup> s<sup>-1</sup>) corresponds to the maximum GPP as  $\text{PPFD} \rightarrow \infty$ , and  $R_{tot}$  is the total ecosystem respiration (mg m<sup>-2</sup> s<sup>-1</sup>). In **Papers III and IV**, the GPP function was extended to explicitly account for the seasonal development in the photosynthesis. This was done with a term called the effective phytomass index ( $PI$ ), developed by Aurela et al. (2001). This was calculated as the difference between the measured average night-time and daytime NEE on a daily time-scale, and was normalized to unity at its annual maximum. Thus,  $PI$  is practically equal to the mean daily GPP. In **Papers I and III**,  $PI$  was also used to account for the seasonal variation in  $R_d$ . The effect of atmospheric vapour pressure deficit on GPP was taken into account during the peak growing season in EC data measured on barley and grass fields (**Paper III**).

To estimate the uncertainty of the CO<sub>2</sub> emission during the growing season, as calculated from the new CO<sub>2</sub> efflux data from bare soil, confidence limits (here the 95% prediction intervals) were estimated using linear regression of the logarithmically-transformed data. The prediction intervals were calculated from the prediction variance  $s_y^2$ , which is defined as:



$$s_y^2 = s_e^2 \left[ 1 + \frac{1}{n} + \frac{(x_0 - \bar{x})^2}{\sum_{i=1}^n (x_i - \bar{x})^2} \right] \quad (6)$$

where  $s_e^2$  is the sample variance of the residuals (obtained from the ANOVA table for linear regression),  $n$  is the sample size,  $x_0$  stands for the soil temperature used for the prediction,  $x_i$  is the soil temperature used for the regression and  $\bar{x}$  is the mean of these temperatures (Wilks 2006). The prediction interval was then added to the predicted soil CO<sub>2</sub> efflux, after which the results were back-transformed. For the wintertime data, the confidence limits were calculated from the standard deviations of the data. The confidence limits for the whole growing season and finally for the whole year were calculated using the standard error propagation method. The difference between the annual cumulative CO<sub>2</sub> emissions was considered statistically significant if the 95% confidence limits did not overlap.

## 5 Results and discussion

### 5.1 Soil respiration in different soil types

#### 5.1.1 Respiration components

The total ecosystem respiration measured by the chamber method was separated into heterotrophic and autotrophic respiration components with a simple temperature response model including photosynthesis (**Paper I**). The partition enabled the comparison of parameters among different soil types. Comparison of the parameter describing soil respiration ( $R_{s0}$ ) at a constant temperature (10 °C) indicated that the potential respiration in the peat soil was 2–3 times as great as that in two different mineral soils (Fig. 3, **Paper I**). With all crop types,  $R_{s0}$  was slightly smaller in the clay than in the sandy soil, which may be linked to the properties of the clay mineral. The mineralization of soil C is retarded by the bonds between the clay particles and organic matter, and on the other hand clay soils favour the formation of macroaggregates, which protect the organic matter from decomposition (Rice 2002). There were also differences in  $R_{s0}$  between the crop types, with the highest rate of  $R_{s0}$  on soil under grass and the lowest on that under potato. This may be explained by the probably higher root biomass of the grass.

The differences in  $R_{s0}$  between the soil and crop types found in this study were supported by direct measurement of soil respiration in different soil types, showing that soil respiration measured at or near 10 °C was always highest in peat soil and in grass (see Fig. 5 in **Paper I**). The one exception was in clay soil in autumn, when respiration was higher in barley than in grass; this can be explained by the 4 °C higher soil temperature.

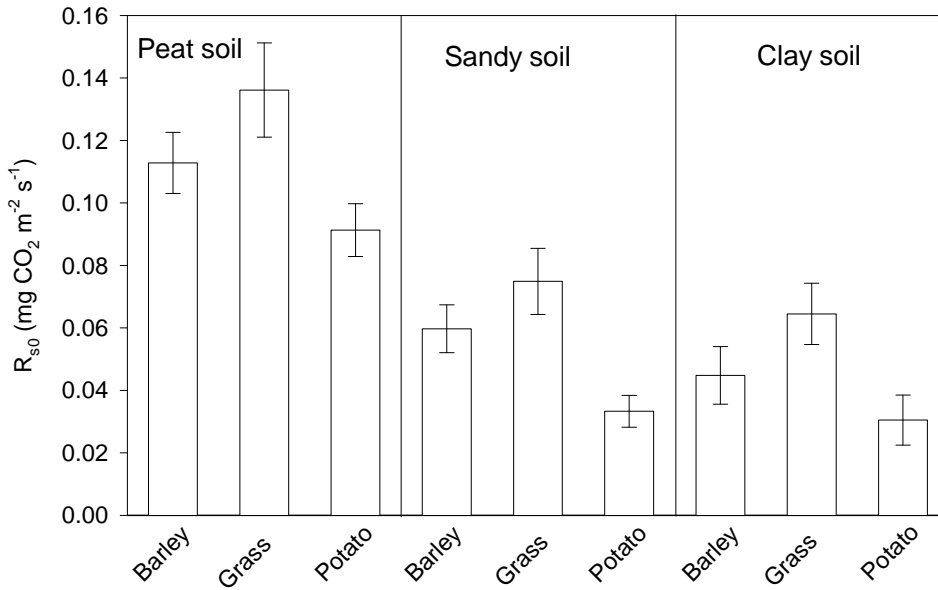


Figure 3. Values of  $R_{s0}$  in different experimental plots at Jokioinen. Error bars denote the 95% confidence limits (redrawn from **Paper I**).

At the height of the growing season, the soil respiration comprised 62%, 46% and 49% of the total ecosystem respiration on the peat, sandy and clay soils, respectively (**Paper I**). These values agree well with those measured on intensively-managed grassland (Byrne and Kiely 2006) and on tallgrass prairie (Franzluebbers et al. 2002), averaging 50% and 40%, respectively. The evolution of the measured total ecosystem respiration closely followed the development of the GPP index ( $PI$ ), suggesting that the photosynthesis-related term should be included routinely as a driving variable when modelling  $R_{tot}$  or  $R_d$ .

It is evident that using the air and soil temperatures as driving variables in the modelling may produce errors in the interpretation of  $R_{s0}$  and  $R_{d0}$ , as these two temperatures may be assumed to correlate. Moreover, the seasonal courses of temperature and plant phenology are definitely correlated, i.e., both reach their maximum in mid-summer. Using empirical models like the one presented here does not give absolute values of  $R_s$  and  $R_d$ , but rather serves as an indicative tool for finding out the relative contributions of the different respiration components.

### 5.1.2 Annual soil CO<sub>2</sub> loss

New data on soil CO<sub>2</sub> efflux measured on bare soil are employed in this thesis in order to calculate the actual CO<sub>2</sub> loss caused by soil respiration. Measured from soil plots with vegetation removed, the rate of CO<sub>2</sub> efflux on the cultivated and afforested peat fields indicated approximately similar soil respiration rates at 3–19 °C (Fig. 4). The temperature response of the soil respiration (parameter  $E_0$ ) was higher on the afforested peat field, but owing to the higher  $R_{s0}$  and soil temperatures on the cultivated peat field, the annual cumulative CO<sub>2</sub> flux there was 1.2 times that at the afforested site (Table 3). However, as

indicated by the 95% prediction intervals, the difference in annual CO<sub>2</sub> emissions between the sites was not statistically significant (Table 3). This is partly due to the substantial scatter in the soil CO<sub>2</sub> efflux data of the cultivated peat field (Fig. 4).

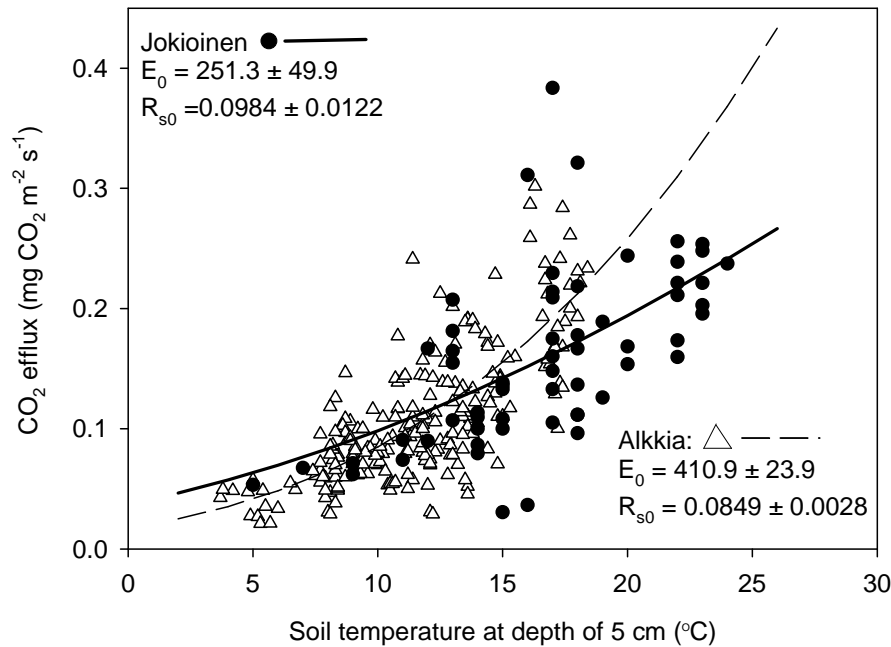


Figure 4. CO<sub>2</sub> efflux from bare soil measured by the chamber method vs. soil temperature at a depth of 5 cm in the Jokioinen agricultural (2002) and Alkkia afforested (2003–2004) peat fields during the growing season. Lines indicate fittings of Eq. (4) to the data. Parameter values  $\pm$  standard errors are also shown (the units of  $E_0$  and  $R_{s0}$  are K and mg CO<sub>2</sub> m<sup>-2</sup> s<sup>-1</sup>, respectively).

It seems that a high rate of peat decomposition continues for a long time after the cultivation of the peat field has finished, the potential for large C loss also being high in the afforested peat field. Interestingly, when compared to the Alkkia afforested peat field, as high and even higher (1900 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup>) rates of annual soil respiration have been measured in nutrient-rich forestry-drained peatlands in both southern and northern Finland (Minkkinen et al. 2007). These study sites have never been under cultivation, and the measurements were conducted with exactly the same methods as those used in the afforested peat field at Alkkia. In more nutrient-poor forestry-drained sites, annual CO<sub>2</sub> losses from the peat were lower, ranging from 900 to 1290 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup> (Minkkinen et al. 2007). The nutrient content of the peat thus seems to play an important role in regulating the greenhouse gas fluxes in peatlands. In this respect, the lower C/N ratio and higher pH, indicating higher fertility, in the cultivated peat field at Jokioinen do not conflict with the higher CO<sub>2</sub> loss there.

Table 3. Annual cumulative CO<sub>2</sub> emission estimates based on chamber measurements on bare soil in various cultivated soils and in afforested peat soil. Emission ranges in parenthesis indicate the 95% prediction intervals.

Site	Measurement period	Annual emission	Winter	Summer	Mean temperature (°C)	
			(October – March)	(April – September)	Winter	Summer
			g CO <sub>2</sub> m <sup>-2</sup> yr <sup>-1</sup>			
Afforested peat soil <sup>a</sup>	7 June 2003 – 6 June 2004	1760 (1350...2290)	240 (140...330)	1520 (1230...1930)	-2.65	11.3
Cultivated peat soil	1 Oct. 2001 – 30 Sept. 2002	2100 (1260...3000)	460 (190...720)	1640 (1110...2430)	-1.35	13.0
Cultivated clay soil	1 Oct. 2001 – 30 Sept. 2002	610 (380...1040)	130 (60...270)	480 (320...730)	„	„
Cultivated sandy soil	1 Oct. 2001 – 30 Sept. 2002	870 (700...1140)	150 (100...230)	720 (490...1070)	„	„

<sup>a</sup>data from Mäkiranta et al. (2007) (except the uncertainty ranges)

The annual CO<sub>2</sub> loss from mineral soils was significantly lower than from peat soils, the emissions of the peat field exceeding those measured on mineral soils by a factor of 2.5–3.5 (Table 3). This peat/mineral soil ratio is approximately the same as the ratio of the modelled  $R_{s0}$  parameters (Fig. 3), supporting the idea that the model works satisfactorily with such data. Similarly to the ratios observed with  $R_{s0}$ , the CO<sub>2</sub> loss from the sandy soil was also higher than that from the clay soil. The results of this exercise indicate that the CO<sub>2</sub> emissions following the cultivation of peat soil are significantly high when compared to agriculture on mineral soils.

## 5.2 Wintertime CO<sub>2</sub> fluxes

### 5.2.1 Temperature effects

During the dormant period, a small CO<sub>2</sub> efflux, measured by EC, was observed on cultivated and afforested peat fields (**Papers II–IV**), with mean rates of 0.016 and 0.036 mg CO<sub>2</sub> m<sup>-2</sup> s<sup>-1</sup>, respectively. On a daily scale, the Jokioinen peat field emitted 0.5–2 g CO<sub>2</sub> m<sup>-2</sup> d<sup>-1</sup> (**Paper III**) while at Alkkia the daily flux was higher, varying mainly between 2 and 4 g CO<sub>2</sub> m<sup>-2</sup> d<sup>-1</sup> (**Paper IV**). The effect of the crop type on the wintertime CO<sub>2</sub> efflux was studied by plotting the weekly medians of the half-hourly NEE against the mean temperature of the 0–50 cm soil layer (Fig. 5). Although the response function (Eq. 4) fitted for individual winters indicate slightly different temperature responses of soil CO<sub>2</sub> efflux, especially for 2001–2002, this difference was not, however, significant. This may be due to the small variation in soil temperatures during this winter.

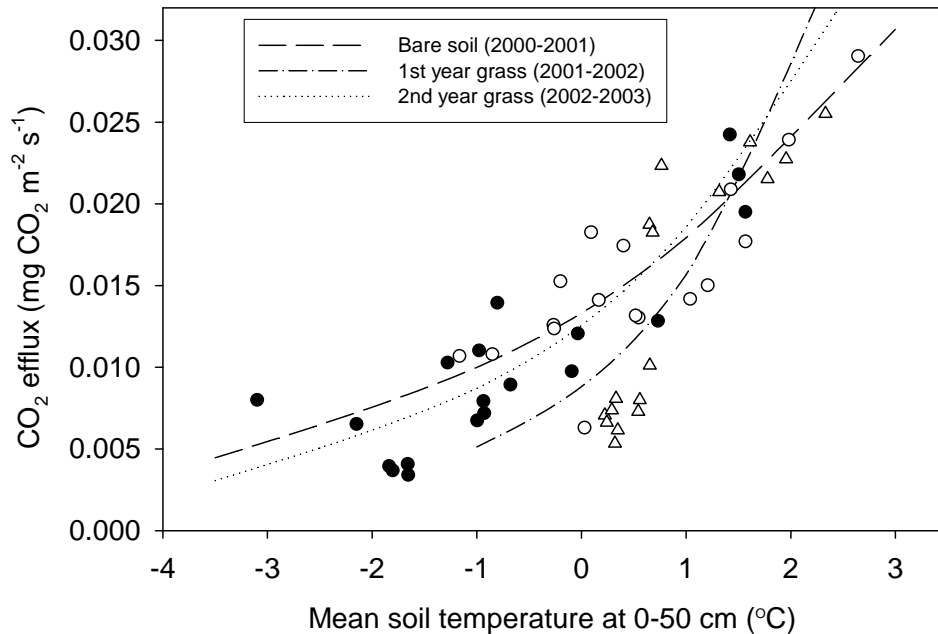


Figure 5. Weekly median fluxes of CO<sub>2</sub> in the Jokioinen agricultural peat field during three winters measured by the EC method against the mean soil temperature of the 0–50 cm soil column. Eq. (4) has been fitted to data for each winter (open circles = bare soil 2000–2001; open triangles = grass 2001–2002; closed circles = grass 2002–2003) (**Paper II**).

At Jokioinen, the rate of CO<sub>2</sub> efflux in winter was positively correlated with the soil temperature (Fig. 5). After the freezing of the soil's upper layers, better correlations were obtained with deeper soil temperatures, indicating that most of the CO<sub>2</sub> production was then taking place in the deeper soil layers (Table 2 in **Paper II**). Air temperature and CO<sub>2</sub> efflux are closely linked at sites like Jokioinen, where soil and air temperatures are rather well coupled due to the typically thin snow layer of less than 30 cm. The thicker snow layer at Alkkia better insulates the soil from the low air temperatures, which favours higher soil respiration rates. In a forest ecosystem, part of the wintertime CO<sub>2</sub> emission results from the respiration of tree needles, branches and trunks, which partly explains the higher respiration at Alkkia.

Seasonal CO<sub>2</sub> emissions during three winters at Jokioinen correlated with mean soil and air temperatures, as well as with the length of the snow-free period (**Paper II**). Based on the temperature correlation, it can be estimated that in the future warmer climate the wintertime CO<sub>2</sub> loss from cultivated peat soils may exceed the current rates by 30–200%, with the projected wintertime temperature rise of 4–11 °C predicted to occur by 2080 (Jylhä et al. 2004). Especially the results obtained with the highest temperatures should be treated with caution, not only because they represent an extreme scenario, but also since these temperatures exceed the range of the measured data, potentially causing large uncertainties in the predictions.

In a warmer climate, the snow cover in southern Finland will be thinner, lessening the insulating effect of snow. Although the soil frost period will shorten, the probability of frozen soil will be higher in the future (Venäläinen et al. 2001). This may decrease the CO<sub>2</sub> efflux and even compensate the effect of higher temperatures. On the other hand, winters with no freezing or snow cover at all will be more common, at least in southern Finland. In such conditions the soil will stay unfrozen and very moist most of the time in the winter. This will, in the light of the results shown above, probably increase the CO<sub>2</sub> loss from the soil due to higher soil temperatures, provided that the excess moisture does not limit decomposition. Measurements should be made in such conditions to get information on the dynamics of soil respiration during much warmer winters.

### 5.2.2 Soil and snow thawing effects

In addition to the longer-term dynamics of the wintertime CO<sub>2</sub> efflux, some short-term variation occurred that was not explained by the soil temperature: changes in the properties of the soil surface during transition stages, like snow or soil thawing, temporarily decreased the rate of CO<sub>2</sub> efflux. These changes were linked to a drop in surface soil albedo, reflecting changes in snow cover and corresponding to the wetness of the surface soil. The depressed rate of CO<sub>2</sub> emission was probably connected to the lowered diffusion of CO<sub>2</sub> from the soil into the atmosphere, attributable to the high value of water-filled pore space after the snow melt (**Paper II**).

In January–February 2002, the depression of the CO<sub>2</sub> efflux rate after a warm spell and partial snow melt occurred roughly at the same time as an increase in soil CO<sub>2</sub> concentration in the 5–25 cm soil layer. However, the concurrent concentration increase was only able to explain about 20% of the lowered efflux.

Soil thawing in spring, on the other hand, induced CO<sub>2</sub> pulses in two years out of three. In 2000–2001 the pulse was relatively small. In 2003, the pulse was more intensive, but the interpretation of the data is more complicated due to a long data gap just before the occurrence of the peak. Moderate spring pulses of CO<sub>2</sub> (Friborg et al. 1997) and CH<sub>4</sub> (Rinne et al. 2007) have been detected earlier in mires with the EC technique; however the importance of these pulses to the annual balance seems to be low. It was suggested that the pulses may arise from a release of gases trapped in frozen soil. In spring 2002, soil CO<sub>2</sub> concentration data available for the grass field indicated a substantial decrease in soil CO<sub>2</sub> concentration during the soil melting, but no pulse was observed then; it may have been masked by the plant respiration of the growing grass.

Snow thawing in mid-winter did not only affect the rate of CO<sub>2</sub> efflux by altering the conditions of the soil-air interspace; in February 2002, shortly after the snow melt, the awakening of grass photosynthesis was observed, provided that the air temperature was above about -1.5 °C. The spring recovery of the grass assimilation after the mild winter of 2001–2002 was controlled by snow cover and air temperature, and after the severe winter of 2002–2003 by soil thawing (**Paper II**).

Although the depression of CO<sub>2</sub> flux induced by snow-thawing is an interesting phenomenon, its contribution to the annual CO<sub>2</sub> balance is minor. However, it should be taken into account in interpreting flux measurements to construct response models. In future, the significance of the mid-winter photosynthesis to the gas balances may be greater in a warmer climate when photosynthesis may potentially continue throughout the winter. Even then, it will be limited by the low radiation levels in northern latitudes. On the other hand, the increasing soil respiration from the warmer soil will probably counterbalance the amount of CO<sub>2</sub> taken up by grass during the winter.

### 5.3 Growing season CO<sub>2</sub> fluxes

The seasonal variability of the CO<sub>2</sub> fluxes was high on both afforested and cultivated peat soils (Fig. 6, **Papers I, III and IV**). Although there were individual days with positive daily NEE, mainly negative daily NEE was observed from May to September in the peat field growing grass (Fig. 6b, **Paper III**) and on the afforested peat field (**Paper IV**). However, in September a daily net CO<sub>2</sub> uptake occurred only during 8 days in grass, and the magnitude of NEE was low. In the peat field growing barley, the period of negative NEE was much shorter still, beginning in late June and ending in early August (Fig. 6a, **Paper III**).

On the cultivated peat field, the seasonal evolution of the respiration and CO<sub>2</sub> uptake by photosynthesis followed that of the LAI with both barley and grass, the highest values being measured typically in July in the barley field and before the harvests in the grass field (**Papers I and III**). At the afforested site, the maximal rates of night-time respiration and net CO<sub>2</sub> uptake were observed in August (**Paper IV**). The highest short-term (30-min) CO<sub>2</sub> uptake rates, equalling  $-1.0 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ , were measured in the barley field and at the afforested site, whereas the nocturnal CO<sub>2</sub> efflux rate was greatest at the afforested site,  $0.60 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ . In the grass field, the maximal rate of CO<sub>2</sub> uptake was lower, reaching  $-0.75 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ , but the canopy respiration rate was similar to that of barley,  $0.35 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ .

The highest rates of daily  $R_{tot}$  were slightly more positive, and those of GPP more negative, in the barley field as compared to the grass (Fig. 6). At the afforested site, the highest positive daily NEE and the highest daily net uptake of CO<sub>2</sub> were similar to those in the barley field, about  $10$  and  $-20 \text{ g CO}_2 \text{ m}^{-2} \text{ d}^{-1}$ , respectively, while in the grass field both peak values were slightly more positive.

Albeit the daytime chamber (**Paper I**) and night-time EC (**Paper III**) data show that in grass the total ecosystem respiration peaked in July before the second harvest, the highest positive daily NEE was, however, observed immediately after both harvests in June and August. GPP was then greatly reduced due to the removal of the biomass, whereas  $R_{tot}$  decreased only slightly owing to the cessation of above-ground plant respiration. As a result, a high positive NEE of up to  $15 \text{ g CO}_2 \text{ m}^{-2} \text{ d}^{-1}$  was observed. At both sites and with both crops, positive values of daily NEE were detected even in the middle of the growing season on cloudy days. While the afforested site acted as a daily CO<sub>2</sub> source only occasionally on a few isolated cloudy days during the growing season, there were longer continuous periods

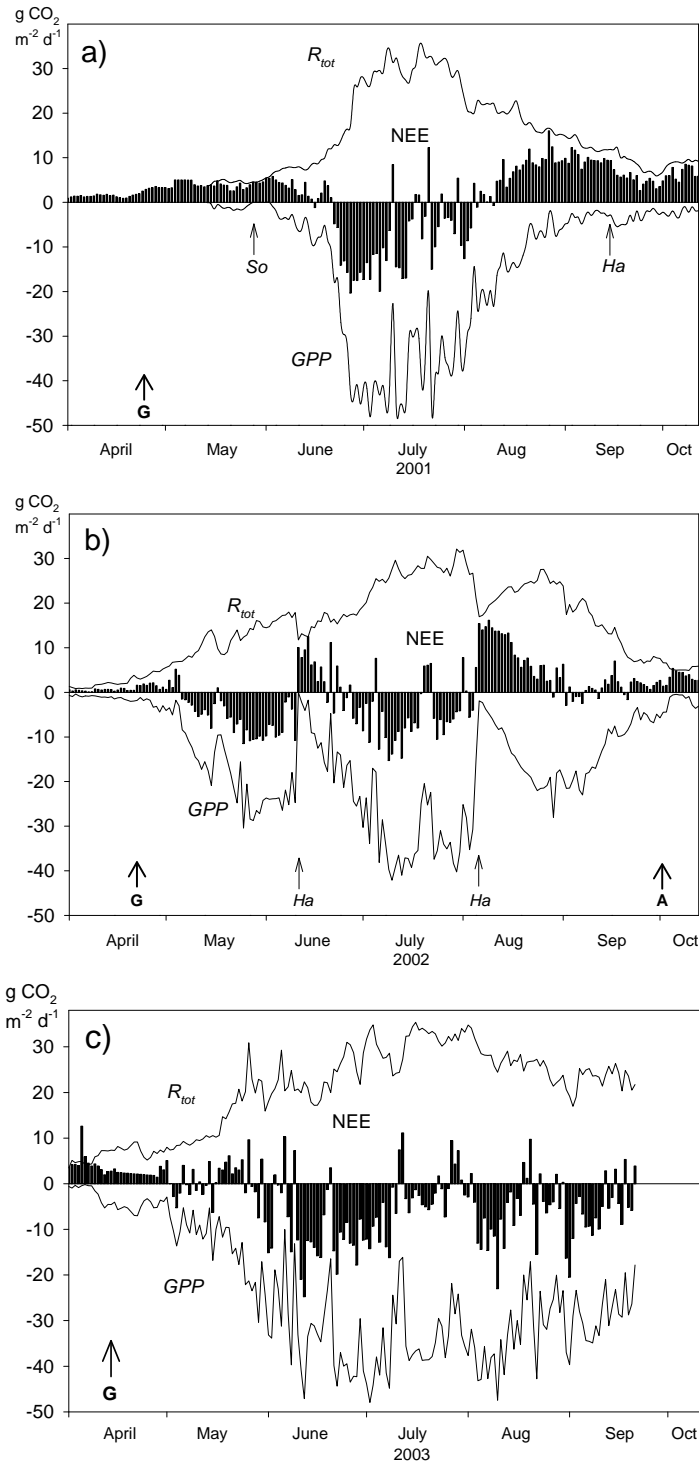


Figure 6. Daily NEE (black bars),  $R_{tot}$  (upper line) and GPP (lower line) from 1 April to 11 October in the Jokoinen peat field growing (a) barley and (b) grass and (c) from 1 April to 21 September in the Alkkia afforested peat field. So = sowing, Ha = harvest; G and A indicate the start of the thermal growing season and autumn, respectively (a and b redrawn from **Paper III**).



of positive daily NEE at the cultivated site due to harvests (grass) or due to the missing green leaf area before and after the period of active growth (barley) (Fig. 6).

## 5.4 Annual balances

### 5.4.1 Net ecosystem CO<sub>2</sub> exchange

The cultivated peat field at Jokioinen was a large source of CO<sub>2</sub> to the atmosphere. A higher positive NEE,  $770 \pm 100 \text{ g CO}_2 \text{ m}^{-2} \text{ yr}^{-1}$  ( $210 \pm 27 \text{ g C m}^{-2} \text{ yr}^{-1}$ ) was observed in the barley field, as compared to the grass with an annual balance of  $290 \pm 90 \text{ g CO}_2 \text{ m}^{-2} \text{ yr}^{-1}$  ( $79 \pm 25 \text{ g C m}^{-2} \text{ yr}^{-1}$ ). The uncertainty values include estimates of the errors considered to be the most important: the random error in the measured fluxes and errors associated with model uncertainty, the scarcity of night-time flux data on summer nights, the high-frequency correction of the EC measurement, the CO<sub>2</sub> analyzer calibration, and the effect of the different vegetation in the adjacent experimental plot (**Paper III**).

The large difference in NEE between years arose mostly from the different crops and hence different management: a large amount of carbon was lost from the barley field in spring after the soil thawing until the barley sowing; in this period there was no vegetation. This is well seen in the growing season CO<sub>2</sub> balances in 2001 and 2002: the grass ecosystem accumulated approximately  $60 \text{ g CO}_2 \text{ m}^{-2}$  during the growing season of 2002, whereas a cumulative loss of about  $360 \text{ g CO}_2 \text{ m}^{-2}$  was observed in the barley field in 2001.

While the mean annual temperatures were practically the same during the study period (Table 1 in **Paper III**), the monthly means differed: November and December were much colder during the first year, whereas February and March were colder in the second year. The annual precipitation in 2000–2001 was 1.43 times that in 2001–2002, with the greatest difference in August and September, which were very dry in 2002. Owing to the wide range of factors affecting the growing season CO<sub>2</sub> exchange on cultivated peatland, e.g., climate, cultivable crop, fertilization, ploughing, etc., it is difficult to determine the exact contribution of the meteorological factors. However, the growing season balances were particularly affected by the crop management and its timing, whereas the wintertime balances were mainly influenced by the meteorology, especially the temperature (**Paper II**).

By contrast to the cultivated organic soil, the afforested peat field was nearly in balance with respect to CO<sub>2</sub>: the annual NEE at Alkkia was  $50 \text{ g CO}_2 \text{ m}^{-2} \text{ yr}^{-1}$  (with an uncertainty range from  $-163$  to  $154 \text{ g CO}_2 \text{ m}^{-2} \text{ yr}^{-1}$ ) (**Paper IV**). This value indicates that the NPP by the trees and the forest floor vegetation roughly balanced the decomposition of the underlying peat layer. Adding the peat decomposition rate of  $1760 \text{ g CO}_2 \text{ m}^{-2} \text{ yr}^{-1}$  (Table 3, Mäkiranta et al. 2007) to the NPP of trees ( $-870 \text{ g CO}_2 \text{ m}^{-2} \text{ yr}^{-1}$ ) and subtracting the annual NEE from this produces an imbalance of about  $-840 \text{ g CO}_2 \text{ m}^{-2} \text{ yr}^{-1}$ . This difference is due to CO<sub>2</sub> flux components that were not measured and accounted for.

First of all, the annual net accumulation of CO<sub>2</sub> into tree biomass, calculated from the tree measurements, was probably a low estimate of the actual tree biomass growth, because fine

roots were not included in the calculation (**Paper IV**). Second, the birch dry mass accumulation may also have been underestimated. Third, a large part of the discrepancy may be explained by the possible accumulation of C by ground vegetation biomass and below-ground litter production (Laiho et al. 2003). Since a considerable amount of carbon may be bound in recently-allocated litter, an exact quantification of the magnitude of these pools and fluxes should be done in future carbon balance studies at drained peatlands. Finally, the employment of different methods (chambers, EC, measurements of growth of trees) in the determination of the forest CO<sub>2</sub> balance introduces some uncertainty into it. For example, the EC method and the chambers measure at different spatial scales.

Figure 7, showing the cumulative CO<sub>2</sub> exchange from the cultivated and afforested peat fields, illustrates the diverse dynamics of the CO<sub>2</sub> balance at different sites and with different plant species. The cumulative wintertime emission was greatest at the afforested site and smallest in grass. The onset of net CO<sub>2</sub> uptake occurred first in grass and later at the afforested site and in barley. However, the cumulative carbon fixation was greatest at the afforested site, the three and half months period of net uptake nearly compensating the preceding emissions during autumn and winter. In barley, a very high emission peak lasting nearly 2 months was observed in the spring and early summer, when the soil was no longer frozen, but there was still no vegetation on the field.

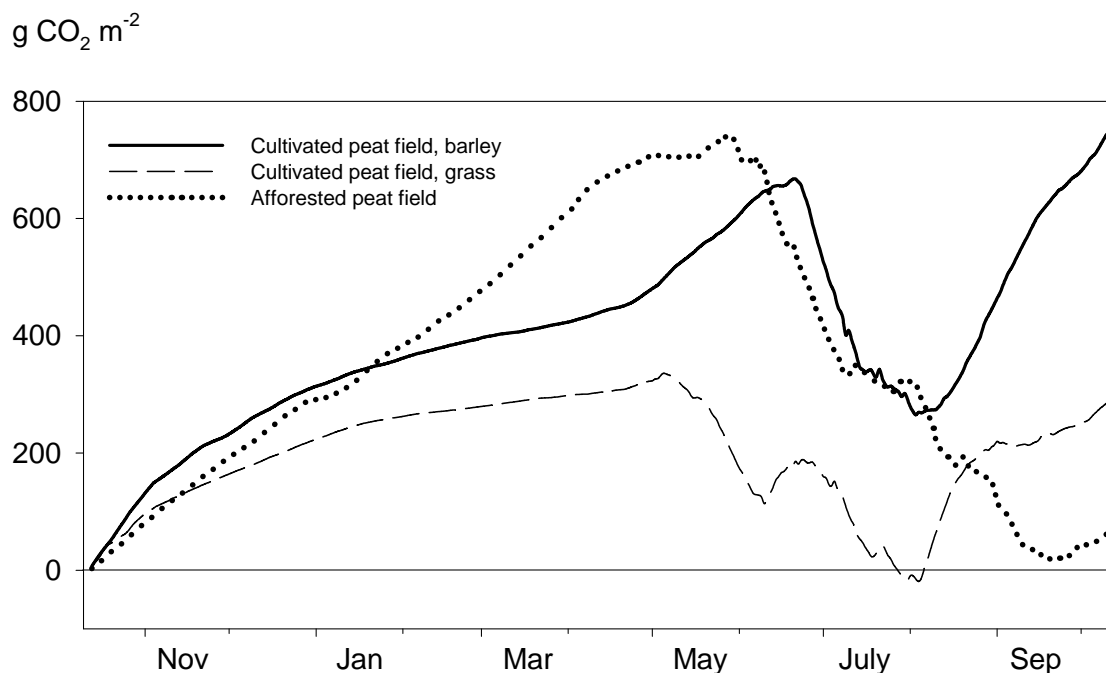


Figure 7. Cumulative annual NEE on the Jokioinen agricultural and Alkkia afforested peat fields.

The duration of the net carbon uptake period (i.e., the season with negative NEE) was clearly longest at the afforested site (110 days). On the barley and grass fields, the net carbon uptake periods were shorter, 40 and 80 days, respectively. This result reflects both

the large CO<sub>2</sub> uptake capability of boreal forests and the strong impact of ecosystem management on CO<sub>2</sub> fluxes. In grass, the CO<sub>2</sub> uptake started as soon as the snow had disappeared and soil thawed, whereas the late sowing of barley delayed the start of the uptake period so that the first day of negative NEE took place as late as the second half of June. Even the slightly higher photosynthetic capacity (**Paper III**) and accordingly greater daily CO<sub>2</sub> uptake in barley during the peak growing season (Fig. 6) was not able to compensate for the C losses from the bare soil in May and June. It can also be speculated that, without harvests, the grass would have soon reached the reproductive growth stage, where leaves start to senesce and photosynthesis is reduced. Therefore, harvesting actually lengthens the period of CO<sub>2</sub> uptake in grass, as it is kept continuously in the growing stage. In addition, the rather similar temperature responses of soil respiration in the cropped and afforested peat fields (Fig. 4), as well as the non-significant difference in annual *in situ* emission from the peat (Table 3) support the idea that the difference in annual NEE is more to be attributed to differences in CO<sub>2</sub> uptake than in soil respiration rate.

Although the net carbon uptake period was considerably longer on the afforested as compared to the cultivated peat field, it was actually short if compared to other boreal forests and to temperate coniferous forests. In Central Europe, the length of the negative NEE season in coniferous forests ranges from 200 to 365 days (Owen et al. 2007). Even in the Hyytiälä Scots pine forest (Suni et al. 2003), located at about the same latitude as Alkkia, the duration of the net carbon uptake period was 166 and 194 days in 2001 and 2002, respectively, about 2–3 months longer than at Alkkia. The length of the negative NEE season in the afforested peat field at Alkkia was comparable to a Scots pine forest at Sodankylä (Aurela 2005, Thum et al. 2007), where the annual temperature is -1.0 °C. The growing season is typically short there, as the site is located north of the Arctic Circle. This comparison highlights the dominant role of the peat decomposition in the ecosystem CO<sub>2</sub> balance.

Year-round EC measurements in European forests have suggested that most of these ecosystems are CO<sub>2</sub> sinks, the sink strength being roughly correlated to latitude with increasing annual sink values towards the south, the annual NEE in coniferous forests ranging from -280 to -2500 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup> (Valentini et al. 2000). At Hyytiälä, the annual NEE varied from -600 to -900 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup> in 1997–2001 (Suni et al. 2003). Studies in afforested peat soils are scarce: there are no direct measurements of total ecosystem CO<sub>2</sub> exchange in such ecosystems. Based on combined measurements and modelling, a Scottish peatland was concluded to become a carbon sink quite soon after its afforestation: 4–8 years after the planting of trees the annual NEE equalled about -1100 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup> (Hargreaves et al. 2003). However, this site had not been in agricultural use.

#### 5.4.2 Carbon balance

To estimate the carbon balance of an agricultural ecosystem, the net biome production (NBP) needs to be calculated by adding the exported biomass to NEE (it is assumed here that all the harvested carbon is ultimately transformed to CO<sub>2</sub>). In contrast to the NEE values, NBPs of the barley and grass fields at Jokioinen were 336 and 452 g C m<sup>-2</sup> yr<sup>-1</sup> respectively, indicating a greater carbon loss from the grass field (**Paper III**). This

calculation does not take into account the fact that for barley more CO<sub>2</sub> was lost directly into the atmosphere, while in the case of grass, the CO<sub>2</sub> loss is partially indirect, the carbon being temporarily bound in biomass that is utilized in cattle husbandry before being emitted into the atmosphere. To assess the climate impact of grass production, the emissions of other GHGs linked to cattle husbandry, such as the CH<sub>4</sub> emissions from ruminants should be taken into account.

Although the afforestation seems to offer an expedient to reduce the short-term CO<sub>2</sub> burden of a cultivated peat soil, carbon loss is probable in the long run, beyond the rotation length. The near-zero NEE (14 g C m<sup>-2</sup> yr<sup>-1</sup>) indicates that the long-term NBP, which was not determined in this study, will be positive, i.e., carbon will be lost from the ecosystem. Then, the mineralization of the old peat will exceed the accumulation of new litter. Overall, our results suggest that the agricultural history has long-lasting effects on the soil properties, sustaining considerable peat decomposition rates. On the other hand, as pointed out in Chapter 5.1.2, the *in situ* soil respiration rates in the afforested site were similar to those in nutrient-rich forestry-drained peatlands, indicating that high CO<sub>2</sub> efflux rates may be found even in peat soils without a farming history.

In this study the leaching and airborne erosion of carbon were not measured; however, it may be assumed that the contribution of these is small as compared to that of the net atmosphere-biosphere CO<sub>2</sub> exchange, at least in the cultivated peat field where the measured carbon loss was more than an order of magnitude higher than the average estimated DOC export by leaching ( $\approx 10$  g C m<sup>-2</sup> yr<sup>-1</sup>). Airborne erosion would probably only be significant in a barley field which was ploughed and stayed vegetation-free for several weeks before sowing. In the afforested site where the annual NEE was small, the carbon lost by leaching probably constitutes a much greater share of the carbon balance. However, including the leaching and airborne erosion would only lead to an even larger carbon loss at every site, and would not change the sign of the annual flux.

The comparison of the NEE of the cultivated peat field at Jokioinen with other peat soil sites indicates that the CO<sub>2</sub> loss was indeed significant there. On a drained peat soil in New Zealand, a relatively small annual CO<sub>2</sub> loss of 16 g m<sup>-2</sup> yr<sup>-1</sup> was observed on a grazed pasture (Nieveen et al. 2005). However, there too the inclusion of other carbon losses (methane emission from cows and milk production) yielded an NBP of 106 g C m<sup>-2</sup> yr<sup>-1</sup>. On a drained peat soil in the Netherlands, a carbon loss equal to our site, about 460 g C m<sup>-2</sup> yr<sup>-1</sup>, was measured at both an intensive dairy farm and in a meadow bird reserve with extensive farming (Veenendaal et al. 2007). At the dairy farm, this estimate also includes the ruminant methane emissions. In a peat field in Eastern Finland drained for agriculture, Nykänen et al. (1995) reported an annual CO<sub>2</sub> flux of 400 g C m<sup>-2</sup> yr<sup>-1</sup> from bare soil, measured by the chamber method. This value is the same as the average measured carbon loss from the barley and grass fields at Jokioinen, and about 70% of the carbon loss measured by chambers in the bare peat field at Jokioinen (Table 3). The average NBP of the Jokioinen peat field is also very close to the mean value summarized by Drösler et al. (2008) for arable peatlands in Europe.

In a mineral agricultural soil growing wheat in Germany, NEE ranged from -680 to -900 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup> while NBP varied from 45 to 105 g C m<sup>-2</sup> yr<sup>-1</sup> (Anthoni et al. 2004a). With potato growing next year at the same site, a smaller annual uptake of CO<sub>2</sub> (NEE = -125 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup>) was observed, becoming a carbon source of 386 g C m<sup>-2</sup> yr<sup>-1</sup> when the harvest was included (Anthoni et al. 2004b). These results suggest that not only peat soils, but also mineral soils under agriculture may be prone to carbon loss, at least in the short term.

## 5.5 Climate effect of land-use

The warming effect of greenhouse gases depends on their radiative efficiency and atmospheric lifetime. These have been taken into account in the concept of global warming potential (GWP), used to compare the climate impact of different GHGs. Generally GWP is expressed as CO<sub>2</sub>-equivalents, meaning that the warming potential of other gases are compared to that of CO<sub>2</sub>. Its magnitude depends on the time-scale adopted: for example, on a 100-year time horizon, CH<sub>4</sub> and N<sub>2</sub>O have 25- and 298-fold efficiencies, respectively, in absorbing radiation as compared to CO<sub>2</sub> (Forster et al. 2007).

In Figure 8, which shows the GHG balances of peatlands under various land-use at certain sites in Finland, different gases have been scaled by the corresponding GWP. Here, the exchange of CO<sub>2</sub> refers to NEE, and the fate and magnitude of biomass carbon has been neglected. At all sites, the annual CO<sub>2</sub> balances have been measured by the EC method. Positive and negative values refer to warming and cooling effects on the atmosphere, respectively. For most of the sites and managements, measurements have only been conducted for 1–1.5 years and, therefore, the interannual variation is mostly unknown. As discussed in Chapter 5.4.1, such variation may be caused by the meteorology or by different management practices, such as a varying litter input after harvest or ploughing methods on a cultivated site. In an afforested site, the time elapsed since thinning and the stage of the rotation, i.e., the age of the trees, have an effect on the gas balances. The variation caused by these factors is not included in this comparison. At the Lompolojänkki and Kaamanen mires, however, the CO<sub>2</sub> flux data shown cover 3 and 6 years, respectively.

Due to the high CO<sub>2</sub> and N<sub>2</sub>O emissions, cropped peat soil growing barley has the largest positive GWP (i.e., potential climate warming effect). Also the grass field and afforested peat field have a positive GWP, but considerably smaller than that of the barley field. In the afforested peat field, the largest warming effect is caused by the high N<sub>2</sub>O emission. Enhanced carbon and nitrogen mineralisation rates are still seen in the peat 30 years after afforestation (**Paper IV**; Mäkiranta et al. 2007). Interestingly, all of the studied undrained mires also have a tendency at present to warm the climate on a 100-year time horizon (Fig. 8). This is mainly due to the methane emissions, which greatly contribute to the net GHG balance on pristine peatlands. However, the use of the GWP methodology for atmosphere-biosphere flux data has been criticized for treating the gas fluxes as pulses instead of sustained and temporally varying emission/uptake rates (Frolking et al. 2006). Using dynamic modelling, Frolking et al. (2006) showed that, after the beginning of their formation, northern mires exert a warming effect on the climate for the first several

hundreds to several thousands of years, the sign of the effect changing to cooling thereafter. This is due to their persistence as a CO<sub>2</sub> sink over a long period.

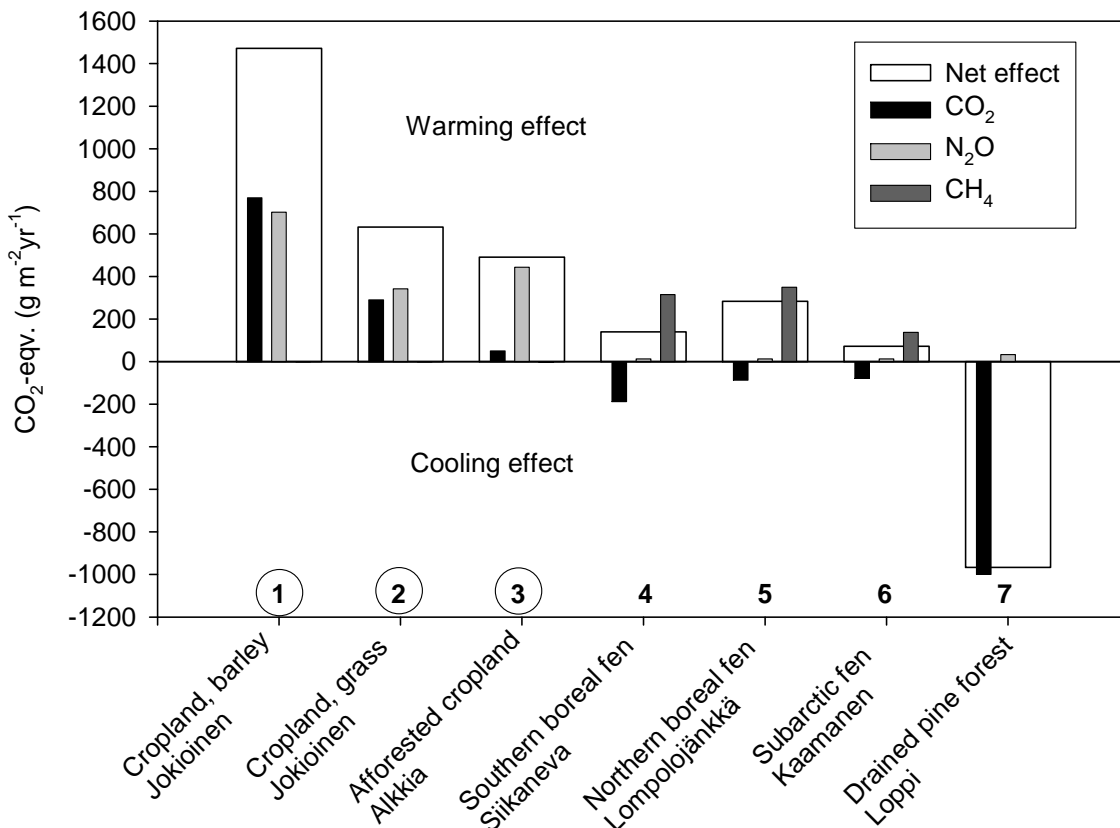


Figure 8. Net GHG balances expressed as CO<sub>2</sub>-equivalents on a 100-year time horizon at pristine (4–6) and drained (1–3, 7) peatlands with different land-use in Finland. Circled numbers below the bars indicate sites where CO<sub>2</sub> data from this study are used. Other data have been collected from (1) and (2) CH<sub>4</sub>: Regina et al. (2007); N<sub>2</sub>O: Regina et al. (2004), (3) Mäkiranta et al. (2007); (4) CO<sub>2</sub>: Aurela et al. (2007a); CH<sub>4</sub>: Rinne et al. (2007); N<sub>2</sub>O: estimated from Lohila et al. (2008), (5) CO<sub>2</sub> and CH<sub>4</sub>: Aurela et al. (2007b); N<sub>2</sub>O: Lohila et al. (2008), (6) CO<sub>2</sub>: Aurela (2005); CH<sub>4</sub>: Hargreaves et al. (2001); N<sub>2</sub>O: estimated from Lohila et al. (2008), (7) unpublished. The figure is based on Laurila et al. (2007).

In this comparison, the only site with a negative GWP (i.e., potential cooling effect) on the atmosphere was a nutrient-poor forestry-drained peatland in southern Finland. There, both CH<sub>4</sub> and N<sub>2</sub>O emissions were fractional, but the site was a considerably large CO<sub>2</sub> sink annually. It needs to be remembered that all of the mires shown in Fig. 8 are nutrient-rich fens, while on the other hand the only forestry-drained peatland is nutrient-poor. Hence, these results can only be generalized to limited types of different peatlands. According to this comparison, forestry on peatlands is the most recommended option when considering the climatic effect of different land-uses. However this comparison clearly shows the need

for further investigation on a wider scale of different peatlands with different nutrient status and climate and during meteorologically different years.

## 6 Summary and conclusions

Measurements conducted on boreal peat soils indicated that the soil respiration in a cultivated peat field was 2–3 times as high as that in cultivated mineral soil types. Direct measurements of soil respiration and the results obtained by separating the autotrophic and heterotrophic components from the measured total ecosystem respiration by modelling suggested this same ratio. The annual cumulative soil respiration in a cultivated peat field was 20% higher than in an afforested peat field, but the difference was not statistically significant. Plant respiration was an important part of the total ecosystem respiration and has a distinct seasonal cycle governed by photosynthesis, which should be taken into account in the modelling of respiration.

The rate of wintertime CO<sub>2</sub> efflux from a cultivated peat soil was mainly controlled by the soil temperature; the rate of CO<sub>2</sub> efflux was not significantly different from bare and grass-covered peat fields. Thawing of snow caused suppression of CO<sub>2</sub> flux in some circumstances, which was possibly linked to the lowered diffusion rates of gases in wet surface soil. Furthermore, there was evidence of pulses of CO<sub>2</sub> efflux during the soil thawing in spring. The number of snow-free days and soil and air temperatures well explained the year-to-year variation in wintertime cumulative CO<sub>2</sub> emission. With the projected temperature increase by the 2080s, the wintertime CO<sub>2</sub> loss from cultivated peat soils would increase by 30–200%, depending on the rate of temperature growth.

Equally high short-term (30-min) CO<sub>2</sub> uptake rates of  $-1.0 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$  were measured during the peak growing season in the barley field and at the afforested site, whereas the nocturnal CO<sub>2</sub> efflux rate was greatest at the afforested site,  $0.60 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ . A daily CO<sub>2</sub> uptake (i.e., negative daily NEE) was observed from May to September in the peat field growing grass and in the afforested peat field, with individual days of net CO<sub>2</sub> emission after harvests or during cloudy weather leading to low levels of radiation. In the peat field growing barley the period of net carbon uptake was much shorter, beginning in late June and ending in early August.

While the day-to-day variation in NEE was linked to the meteorological conditions, the agricultural management strongly affected the seasonal and annual C balance by determining the length of the growing season. As compared to other ecosystems, the seasons of net carbon uptake were short at all of the studied sites, lengthening in the order annual crop < perennial crop < afforested peat field. On an annual basis, cultivated peat soil was a large source of CO<sub>2</sub>. The annual NEE was higher in barley than in grass, equalling 770 and 290 g CO<sub>2</sub> m<sup>-2</sup> yr<sup>-1</sup>, respectively. When the exported biomass was also included, the carbon loss was greater from grass, with NBPs of 340 and 450 g C m<sup>-2</sup> yr<sup>-1</sup> in barley and grass, respectively. The carbon loss by leaching or airborne erosion was not measured, but was considered small with respect to the annual NBPs.

CO<sub>2</sub> emissions from the peat decomposition in the afforested peat field were nearly balanced by the large GPP of the trees and the forest floor vegetation, resulting in an annual NEE of 14 g C m<sup>-2</sup> yr<sup>-1</sup>. Accounting for the whole rotation length will, however, indicate much larger CO<sub>2</sub> loss (i.e., positive NBP) there. Although the CO<sub>2</sub> balance in an afforested organic-soil cropland is close to zero 30 years after the ceasing of cultivation, the net loss of peat still continues.

The GHG emissions of peatlands with different land-uses were compared using the GWP methodology. Nutrient-poor forestry-drained peatlands seem to have a large climate-cooling potential. Owing to their high methane emissions, undrained fens have a tendency to warm the climate on a 100-year time horizon. Cultivated peatlands have a strong climate-warming effect, particularly with annual crops. The net warming effect is more than double that of a peat field growing grass and of an afforested peat field. The results of this study indicate that the high CO<sub>2</sub> losses from agricultural peat soils may be partially moderated by growing perennial crops with a longer growing season or by afforestation, but that these actions do not stop the carbon loss from the peat.

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