



Title	Effect of land use change and drainage on peat decomposition and greenhouse gas emission in a tropical peatland
Author(s)	Adji, Fengky Florante
Citation	北海道大学. 博士(農学) 甲第11395号
Issue Date	2014-03-25
DOI	10.14943/doctoral.k11395
Doc URL	http://hdl.handle.net/2115/56102
Type	theses (doctoral)
File Information	Fengky_Florante_Adji.pdf



[Instructions for use](#)

**Effect of land use change and drainage
on peat decomposition and greenhouse
gas emission in a tropical peatland
(熱帯泥炭地における泥炭分解
および温室効果ガス放出に及ぼす
土地利用変化と排水の影響)**

**Laboratory of Soil Science
Division of Environmental Resources
Graduate School of Agriculture
Hokkaido University
Sapporo - Japan**

FENGKY FLORANTE ADJI

ABSTRACT

To evaluate the hypotheses that 1) plant-mediated oxygen supply should oxidize methane (CH_4) produced in saturated tropical peat soil, and 2) this CH_4 oxidation should reduce total global warming potential (GWP) in a tropical peatland, the author compared the fluxes and dissolved concentrations of greenhouse gases [GHGs; CH_4 , carbon dioxide (CO_2), and nitrous oxide (N_2O)] and dissolved oxygen (DO) at multiple peatland ecosystems in Central Kalimantan, Indonesia. Study ecosystems included tropical peat swamp forest and degraded peatland areas that were burned and/or drained during the rainy season.

CH_4 fluxes were significantly influenced by land use and drainage, which were highest in the flooded burnt sites ($5.75 \pm 6.66 \text{ mgC m}^{-2} \text{ h}^{-1}$) followed by the flooded forest sites ($1.37 \pm 2.03 \text{ mgC m}^{-2} \text{ h}^{-1}$), the drained burnt site ($0.220 \pm 0.143 \text{ mgC m}^{-2} \text{ h}^{-1}$), and the drained forest site ($0.0084 \pm 0.0321 \text{ mgC m}^{-2} \text{ h}^{-1}$). Dissolved CH_4 concentrations were also significantly affected by land use and drainage, which were highest in the flooded burnt sites ($124 \pm 84 \text{ } \mu\text{mol L}^{-1}$) followed by the drained burnt site ($45.2 \pm 29.8 \text{ } \mu\text{mol L}^{-1}$), the flooded forest sites ($1.15 \pm 1.38 \text{ } \mu\text{mol L}^{-1}$), and the drained forest site ($0.860 \pm 0.819 \text{ } \mu\text{mol L}^{-1}$). DO concentrations were influenced by land use only, which were significantly higher in the forest sites ($6.9 \pm 5.6 \text{ } \mu\text{mol L}^{-1}$) compared to the burnt sites ($4.0 \pm 2.9 \text{ } \mu\text{mol L}^{-1}$). These results suggested that CH_4 produced in the peat might be oxidized by plant-mediated oxygen supply in the forest sites. CO_2 fluxes were significantly higher in the drained forest site ($340 \pm 250 \text{ mgC m}^{-2} \text{ h}^{-1}$ with the water table level of -20 to -60 cm) than in the drained burnt site ($108 \pm 115 \text{ mgC m}^{-2} \text{ h}^{-1}$ with the water table level of -15 to $+10$ cm). Dissolved CO_2 concentrations were 0.6 to 3.5 mmol L^{-1} , also highest in the drained forest site. These results suggested the enhanced CO_2 emission by aerobic peat decomposition and plant respiration in the drained forest site. N_2O fluxes ranged from -2.4 to $-8.7 \text{ } \mu\text{gN m}^{-2} \text{ h}^{-1}$ in the flooded sites and from 3.4 to $8.1 \text{ } \mu\text{gN m}^{-2} \text{ h}^{-1}$ in the drained sites. The negative N_2O fluxes might be caused by N_2O consumption by denitrification under flooded conditions. Dissolved N_2O concentrations were 0.005 to $0.22 \text{ } \mu\text{mol L}^{-1}$ but occurred at $<0.01 \text{ } \mu\text{mol L}^{-1}$ in most cases. GWP was mainly determined by CO_2 flux, with highest levels in the drained forest site.

The first hypothesis was supported given that 1) CH_4 emissions in the flooded burnt sites were significantly larger than those in the flooded forest sites, 2) dissolved CH_4 concentrations in the burnt sites were much higher than those in the forest sites, and 3) DO concentrations in the forest sites were significantly higher than those in the burnt sites and which was contributed on CH_4 oxidation under water saturated condition. The CH_4 flux and the dissolved CH_4 concentration at a depth of 20 cm in the drained burnt site were similar to values observed in the forest sites, suggesting that CH_4 oxidation in the surface soil layer occurred. The second hypothesis was weakly supported by the observed GWPs in the flooded burnt sites, which were 20% higher than those in the flooded forest sites. In this study, however, GWP was mainly determined by CO_2 flux. Consequently, GWP and CO_2 flux in the drained forest site were the highest for all study sites, and N_2O flux made little contribution to GWP.

Key words: Dissolved oxygen (DO), greenhouse gas (GHG), Indonesia, tropical peatland, and water table level.

LIST OF CONTENTS

LIST OF CONTENTS	i
LIST OF TABLES	iv
LIST OF FIGURES	v
ACKNOWLEDGEMENTS	viii
I. GENERAL INTRODUCTIONS	1
II. LITERATURE REVIEW	4
2.1 The greenhouse effect and climate change	4
2.2 Tropical peatlands and their characteristics	5
2.2.1 Distribution of tropical peatlands	5
2.2.2 Tropical peatlands under natural conditions	6
2.2.3 C storage and greenhouse gas balance in tropical peatlands	8
2.2.4 Development of tropical peatlands for agriculture	9
2.2.5 Impacts of peatland drainage	11
2.3 Soil CH ₄ flux	13
2.3.1 Magnitude and variations of soil CH ₄ flux	13
2.3.2 Soils as a source or a sink of atmospheric CH ₄	14
2.3.3 Environmental factors controlling CH ₄ flux	16
2.3.3.1 Land use	16
2.3.3.2 Drainage	18
2.3.3.3 Other environmental factors	19
2.4 Soil CO ₂ flux	19
2.4.1 Magnitude and variations of soil CO ₂ flux	20
2.4.2 Soil CO ₂ flux and the ecosystems' C balance	21
2.4.3 Environmental factors controlling soil CO ₂ flux	22
2.4.4 Autotrophic and heterotrophic respirations in soils	22
2.5 Soil N ₂ O flux	23
2.5.1 Magnitude of soil N ₂ O flux under different climate	24
2.5.2 Temporal and spatial variation in soil N ₂ O flux	25
2.5.3 Environmental factors controlling N ₂ O flux	27
2.5.3.1 Abiotic factors	27
2.5.3.2 Biotic factors and human-induced influences	30
2.5.4 Nitrification and denitrification for N ₂ O dynamics	31
2.6 Global warming potential (GWP)	33
2.7 GHG concentrations dissolved in peat pore water	34
2.8 DO concentration in peat pore water	36
III. DESCRIPTIONS OF STUDY SITE	44
3.1 Overview of Central Kalimantan	44
3.2 Study site	44

IV. ENVIRONMENTAL FACTORS ON TROPICAL PEAT DECOMPOSITION AND GHGs DYNAMICS	47
4.1 Introduction	47
4.2 Materials and methods	47
4.2.1 Precipitation and water table level	47
4.2.2 Air and soil temperatures	48
4.2.3 DO concentration	48
4.2.4 pH and EC	50
4.2.5 Dissolved nitrogenous ions	50
4.2.6 Statistical analysis	51
4.3 Results and discussion	51
4.3.1 Precipitation and water table level	51
4.3.2 Air and soil temperatures	52
4.3.3 DO concentration	53
4.3.4 pH and EC	57
4.3.5 Dissolved nitrogenous ions	58
4.4 Summary	61
V. DYNAMICS OF GHG FLUXES UNDER DIFFERENT LAND USE AND WATER TABLE LEVELS	62
5.1 Introduction	62
5.2 Materials and methods	62
5.2.1 Measurement of GHG fluxes	62
5.2.2 GWP calculation	63
5.2.3 Statistical analysis	63
5.3 Results	64
5.3.1 Soil CH ₄ flux	64
5.3.2 Soil CO ₂ flux	66
5.3.3 Soil N ₂ O flux	68
5.3.4 GWP	72
5.4 Discussion	73
5.4.1 Dynamics of soil CH ₄ flux in tropical peatlands	73
5.4.1.1 Effect of water table level	73
5.4.1.2 Impact of land use change	74
5.4.2 Dynamics of soil CO ₂ flux in tropical peatlands	75
5.4.2.1 Effect of water table level	75
5.4.2.2 Impact of land use change	76
5.4.3 Dynamics of soil N ₂ O flux in tropical peatlands	77
5.4.3.1 Effect of water table level	77
5.4.3.2 Impact of land use change	78
5.4.4 Effect of each GHG flux on GWP	79
5.5 Summary	79
VI. DYNAMICS OF DISSOLVED GHG CONCENTRATIONS UNDER DIFFERENT LAND USE AND WATER TABLE LEVELS	80
6.1 Introduction	80

6.2	Materials and methods	80
6.2.1	Concentration of dissolved CH ₄ and N ₂ O	80
6.2.2	Concentration of dissolved CO ₂	81
6.2.3	Calculation of dissolved GHG concentrations	82
6.2.4	Statistical analysis	83
6.3	Results	84
6.3.1	Dissolved CH ₄ concentration	84
6.3.2	Dissolved CO ₂ concentration	86
6.3.3	Dissolved N ₂ O concentration	88
6.4	Discussion	91
6.4.1	Concentrations of dissolved CH ₄ in tropical peatlands .	91
6.4.2	Concentrations of dissolved CO ₂ in tropical peatlands .	92
6.4.3	Concentrations of dissolved N ₂ O in tropical peatlands .	93
6.5	Summary	94
VII.	GENERAL DISCUSSION	95
7.1	Natural peat swamp forest and their contributions on C balance .	95
7.2	GHG fluxes from peat swamp forest affected by artificial drainage	97
7.3	GHG fluxes from degraded peatland affected by deforestation and fire	98
7.4	Effect of land use change and drainage conditions on C balance in the peatlands	100
VIII.	CONCLUSIONS	101
REFERENCES	103

LIST OF TABLES

<i>No.</i>	<i>Table caption</i>	<i>Page</i>
2.1	Comparison of the fluxes and dissolved concentrations of CO ₂ , CH ₄ , and N ₂ O in wetland ecosystems in the world	39
4.1	Air temperature at all study sites averaged for the observation period (Dec. 2011 to May 2012)	53
4.2	DO concentration under different land use and drainage conditions averaged for the observation period	56
4.3	Results of the three-way ANOVA on DO concentrations among different land use, drainage, and depth	56
4.4	pH and EC at all study sites averaged for the observation period (Dec. 2011 to May 2012)	58
4.5	NO ₃ ⁻ and NH ₄ ⁺ concentrations at all study sites averaged for the observation period (Dec. 2011 to May 2012)	60
5.1	CH ₄ flux measured at the ground surface under different land use and drainage conditions averaged for the observation period	66
5.2	CO ₂ flux measured at the ground surface under different land use and drainage conditions averaged for the observation period	68
5.3	N ₂ O flux measured at the ground surface under different land use and drainage conditions averaged for the observation period	70
5.4	Summary of ANOVA on GHG fluxes and GWP under different land use and drainage conditions and multiple comparison among the study sites	71
5.5	GWP eq. to CO ₂ -C fluxes measured at the ground surface under different land use and drainage conditions averaged for the observation period	72
6.1	Dissolved CH ₄ concentrations under different land use and drainage conditions averaged for the observation period	85
6.2	Dissolved CO ₂ concentrations under different land use and drainage conditions averaged for the observation period	88
6.3	Dissolved N ₂ O concentrations under different land use and drainage conditions averaged for the observation period	90
6.4	Results of three-way ANOVA on the dissolved GHG concentrations among different land use, drainage, and depth	90

LIST OF FIGURES

<i>No.</i>	<i>Figure caption</i>	<i>Page</i>
2.1	Change in abundance of major GHGs for the last thousand years (IPCC, 2007)	4
2.2	Map of peatland distribution in the Southeast Asia (Page <i>et al.</i> , 2004)	6
2.3	Overview of peat swamp forest canopy from the top of a tower established in Sebangau area, Central Kalimantan - Indonesia (Image: Adji F.F)	8
2.4	Schematic representation of C cycle processes, flow paths, and stores in tropical peatland (after Jauhiainen <i>et al.</i> , 2001)	9
2.5	Schematic illustration of drainage effects on a peatland dome (modified from Delft Hydraulics, 2006)	12
2.6	Schemes of C fluxes and factors controlling them in the ecosystems ...	21
2.7	Diagram of the “hole-in-the-pipe” conceptual model describing soil emissions of N ₂ O and NO (Davidson <i>et al.</i> , 2000)	33
3.1	Locations of field observation sites	45
4.1	View of the tipping bucket rain gauge settled near the base camp of Kalampangan canal	48
4.2	View of DO measurement at the study site	50
4.3	Diagram of an in-situ DO measurement	50
4.4	View of a set of water samplers for pH, EC, NO ₃ ⁻ , NH ₄ ⁺ , and DO measurements at the study site	51
4.5	Seasonal changes in precipitation (bar) and water table level at each site (line with symbol)	52
4.6	Seasonal variations in air temperature at each study sites	53
4.7	Vertical distributions in soil temperature at all soil depths. Error bars show standard deviation. For better identifiability, the depths in the profiles were slightly shifted from their actual depths (5, 10, 30, and 50 cm)	54
4.8	Vertical distributions of DO concentration in subsurface pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation. For better identifiability, the depths in the profiles were slightly shifted from their actual depths (10, 20, 40, and 80 cm)	55

4.9	Seasonal variations of DO concentration in subsurface pore water averaged for depths of 10–80 cm at each site. Error bars show standard deviation	55
4.10	Seasonal variations of pH in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation	57
4.11	Seasonal variations of EC in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation	58
4.12	Seasonal variations of NO_3^- concentration in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation (positive side only)	59
4.13	Seasonal variations of NH_4^+ concentration in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation (positive side only)	59
5.1	Schematic diagram of a closed chamber (Toma and Hatano, 2007)	63
5.2	Seasonal variations of CH_4 fluxes measure at the ground surface in the study sites. Error bars show standard deviation	65
5.3	Relationship between CH_4 flux and water table level averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation	65
5.4	Seasonal variations of CO_2 fluxes measure at the ground surface in the study sites. Error bars show standard deviation	67
5.5	Relationship between CO_2 flux and water table level averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation	67
5.6	Seasonal variations of N_2O fluxes measure at the ground surface in the study sites. Error bars show standard deviation	69
5.7	Relationship between N_2O flux and water table level averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation	69
5.8	GWP equivalent to $\text{CO}_2\text{-C}$ flux at the ground surface averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation	72
6.1	View of the rapid equilibration method for the measurement of dissolved CH_4 and N_2O concentrations	81
6.2	View of field settings for dissolved CO_2 concentrations measurement at the study site	82
6.3	Diagram of an in-situ equilibration chamber in the field	82

6.4	Seasonal variations of dissolved CH ₄ concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation	85
6.5	Vertical distributions of dissolved CH ₄ concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation	86
6.6	Seasonal variations of dissolved CO ₂ concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation	87
6.7	Vertical distributions of dissolved CO ₂ concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation	87
6.8	Seasonal variations of dissolved N ₂ O concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation	89
6.9	Vertical distributions of dissolved N ₂ O concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation	89

Acknowledgements

Throughout the course of this thesis, much assistance, advice, and encouragements as given by a number of individuals and personnel, without which the study would not have been possible to be carried out and completed. My thanks are due to the following.

The author wish to express my heartfelt and deepest gratitude, and immense indebtedness to venerable supervisor, Prof. Ryusuke Hatano, whose scholastic, infinite enthusiasm, and exquisite instructions enable me to complete this research, and more importantly, kindle my academic interests in soil science and global environmental change, especially topics of the tropical peatlands.

My sincere gratitude should be extended to Prof. Munehide Ishiguro and Prof. Takashi Hirano, the members of the co-advisory committee, for their guidance, encouragement, invaluable and constructive comments and suggestions on this manuscript.

Dr. Hamada Yohei, the author in particularly wish to thank for giving continued support and help throughout this period of research.

The greatly and many friends in CIMTROP-UNPAR (Ube Tito, Jeni Ricardo, Patih Rumbih, Trianson Rogath, Kitso Kusin, Talau, Chris, Adi, Aliansah, Radius, Pandu, Ikin, and Triana) for kindness help during my research in Palangka Raya, Central Kalimantan, Indonesia.

Prof. Dr. Ir. Y. Sulistiyanto, MP (the Head of Analytical Laboratory, University of Palangka Raya), who has issued the permission borrowing the N₂ tank. Many thanks goes to the staff of Analytical Laboratory, University of Palangka Raya (Lidia and Manyang), who have been helpful throughout the study.

Dr. Ir. Suwido H. Limin, MS (the Head of CIMTROP-UNPAR), who has given the permission and support during my research in Palangka Raya.

Dr. Kanta Kuramochi, many thanks for his encouragement, helpful, and suggestions during my study in Japan.

The Government of Republic of Indonesia (Ministry of National Education), Rector of University of Palangka Raya, Dean of Faculty of Agriculture, University of Palangka Raya, Head of Department of Agriculture, Faculty of Agriculture, University of Palangka Raya who has given the recommendation for postgraduate program in Hokkaido University, Sapporo, Japan.

Prof. Dr. Ir. Salampak Dohong, MS., Prof. Dr. Ir. Sulmin Gumiri, M.Sc., and Prof. Dr. Ir. Bambang Hendro Sunarminto, SU., who has given the permission and recommendation for this study.

The ex-secretary Ms. Shiori Asano and the present secretary Mrs. Yoshino Nakamura, also Ms. Masumi Kinoshita (student affairs) the author deeply appreciated for their help in dealing with various official matters related.

The Provincial Government of Central Kalimantan (Governor of Central Kalimantan Provincial) for contributed in the financial support for this study.

The President Hokkaido University Fellowship and GCOE-Project support unit for supported in financial to my study in Japan and Indonesia.

All colleagues in the Soil Science Laboratory, Faculty of Agriculture, Hokkaido University for their kindness, support, and encouragement during study and life in Japan.

Many thanks to Mr. Untung Darung, Mr. Haiki Mart Yupi, Mr. Hendrik Segah Patianom and family, Mr. Rony Teguh and family, Mr. Atfritedy Limin, Ms. Nina Yulianti, Ms. Febrina Natalia Bahat, Mrs. Yulinda Wulandari, Mrs. Mandagi Ixchel Feibie and family, and also PPI H/S friends in Sapporo for their kindness and support.

Finally, my special thanks to beloved parents: Florante K. Adji and Herliany D. Gentong, parents-in-law: Matheus Ngadijono and (RIP) Herwaty HS. Tumon, Fredyantoni F. Adji (brother), Agus Hertanto and Guntoro Purwo Handoko (brother-in-law), Rina Mariana, Susi Dwi Hartanti, and Imelda (sister-in-law), also Tama, Nata, Gita (nephews), Santy Marlina (wife), Danika Ziva Angela (daughter), and Darren Floriartha Adji (son), for their understanding, support, and love that make my study abroad go on well.

GENERAL INTRODUCTIONS

Tropical peatland ecosystems in Southeast Asia cover 24.8 million hectares, comprising 56% of the tropical and 6% of the global peatland area. This ecosystem's high carbon (C) density allows for a large regional peat C store of 68.5 Pg, equivalent to 77% of the tropical and 11–14% of the global peat C store (Page *et al.*, 2011a), which is comparable to the total fossil fuel emissions released throughout the world over 9 years (van der Werf *et al.*, 2008). Given the size of this C reservoir, tropical peatlands play a critical role as a global sink for gaseous C. Inversely, if environmental conditions (e.g., land use type, groundwater level) change, these peatlands may become a major source of gaseous C and greenhouse gases (GHGs), such as carbon dioxide (CO₂), nitrous oxide (N₂O) and methane (CH₄), and therefore may increase their global warming potential (GWP).

In recent decades, considerable portions of peatland in Southeast Asia have been reclaimed to support agroforestry and rice paddy development. The deforestation and development of peatlands is usually accompanied by the draining of water from these areas, releasing much C through aerobic peat decompositions. In addition, the water table level can drastically decrease during exceptionally long dry periods, which frequently occur in El Niño and Southern Oscillation (ENSO) years (Hirano *et al.*, 2007 and 2012; van der Werf *et al.*, 2008). This has increased large scale peat fire in Southeast Asia. According to van der Werf *et al.* (2008), the averaged annual C emission by peat fire during 2000–2006 from equatorial Asian countries (Indonesia, Malaysia, and Papua New Guinea) was 0.128 ± 0.051 PgC yr⁻¹, most of them were attributed to Kalimantan and Sumatra Islands, which was comparable to emissions from fossil fuel combustion in these countries (0.148 PgC yr⁻¹ during 2000–2004).

Several trials have been conducted to restore lowland tropical peatlands (Page *et al.*, 2009). For example, restoration efforts have included raising and stabilizing the water table level by blocking drainage canals, which reduces aerobic peat decomposition in the flooded peatlands. However, flooding can also increase anaerobic CH₄ production. Soil denitrification, which is a major contributor to N₂O production, is also generally promoted under reduced conditions. The radiative forcing effects of CH₄ and N₂O are 25 and 298 times higher, respectively, than that of CO₂ (over a 100-yr time horizon; IPCC, 2007). Consequently, the increases in the production of CH₄ and N₂O under flooded conditions could possibly offset decreases in aerobic peat decomposition and associated CO₂ emissions. A large portion of previously drained peatland has lost its original vegetation and surface

layer through deforestation and/or fire. Under such conditions, the recovery of the water table level will not be able to restore ecosystem function and C balance to those associated with original vegetation levels. The change in GHG dynamics in tropical peatland soils after drainage, deforestation, peat fire, and re-wetting has not been fully studied to date.

After reviewing the existing data, Couwenberg *et al.* (2010) have demonstrated that the range of CH₄ fluxes in tropical peatlands was one-fifth that observed in temperate and boreal peatlands. They attributed these low CH₄ emissions to the poor quality of woody tropical peat, which contains higher levels of recalcitrant materials (e.g., lignin). Williams and Yavitt (2010) reported that the biochemical compositions of lignin affect soil methanogenesis. Another explanation involved the oxygen supply from plant roots. Vascular plants in wetlands usually possess special morphological adaptations, such as hypertrophied lenticels, adventitious roots and enlarged aerenchyma (Joabsson *et al.*, 1999), which promote gas exchange between the atmosphere and the rhizosphere and allow for the entry of oxygen to the root zone (Kozłowski, 1997; Megonigal and Day, 1992). According to a recent review by Laanbroek (2010), the percentage of CH₄ oxidized before entering the atmosphere ranges from 0 to up to >90% of the potential CH₄ efflux. Several studies conducted in temperate and boreal wetlands also reported that an increase in measurement depth was associated with a decrease in the amount of root tissue and dissolved oxygen (DO) concentrations and an increase in dissolved CH₄ concentrations (Fritz *et al.*, 2011; Liebner *et al.*, 2012; van der Nat and Middelburg, 1998). Thus, if plant-mediated oxygen transport and CH₄ oxidation also occur in tropical peatlands, significant differences in soil CH₄ fluxes, dissolved CH₄ concentrations, and DO concentrations in peat pore water would be expected between vegetated and non-vegetated lands.

In the relationships of DO and GHGs dynamics, especially CH₄ dynamics at the vegetated and non-vegetated lands in tropical peatland, the following hypotheses are addressed: 1) DO supplied from plant roots should oxidize CH₄ produced in saturated tropical peat soil, and 2) this CH₄ oxidation should reduce total GWP in a tropical peat swamp forest. In order to evaluate these hypotheses, DO and the fluxes and dissolved concentrations of GHGs were measured during the rainy season and compared between at natural forests and burnt area located in a tropical peatland near Palangka Raya, Central Kalimantan, Indonesia under drained and undrained conditions.

This thesis is composed of 8 chapters. The current chapter of Chapter 1 introduces the background and objectives of this study. Chapter 2 provides a relevant review of the literature of tropical peatlands, fluxes and dissolved concentrations of CH₄, CO₂, and N₂O, and DO.

Chapter 3 describes the study sites. Chapter 4 shows the environmental factors affecting the dynamics of GHGs and DO, including materials and methods. Chapter 5 shows the results and discussion about the fluxes of CH₄, CO₂, and N₂O under different land uses and drainage conditions, and their relationship with GWP, including materials and methods. Chapter 6 shows the results and discussion about the dissolved concentrations of CH₄, CO₂, and N₂O under different land uses and drainage conditions, including materials and methods. Chapter 7 describes general discussion. Chapter 8 shows the conclusions of this study as a summary of the entire study.

LITERATURE REVIEW

2.1 Greenhouse effect and climate change

Greenhouse effect is a process by which thermal radiation from a planetary surface is absorbed by atmospheric GHGs, and is re-radiated in all directions. Since part of this re-radiation is back towards the surface, energy is transferred to the surface and the lower atmosphere. As a result, the temperature is higher than it would be if direct heating by solar radiation were the only warming mechanism. Without this effect the Earth's surface would be 33°C colder, i.e. the mean global temperature would be -18°C instead of the current 15°C (Houghton *et al.*, 1995). Thus, the greenhouse effect is essential for the life forms that have developed on Earth. Global climate change is one of the most important issues of contemporary environmental safety. Since the mid 1800s, however, the concentrations of GHGs in the atmosphere have been increasing due to human activities (IPCC, 2001), promoting further global warming. It is generally accepted that the gradual increase in the mean temperature of the Earth's surface is primarily due to this rising GHG concentrations in the atmosphere.

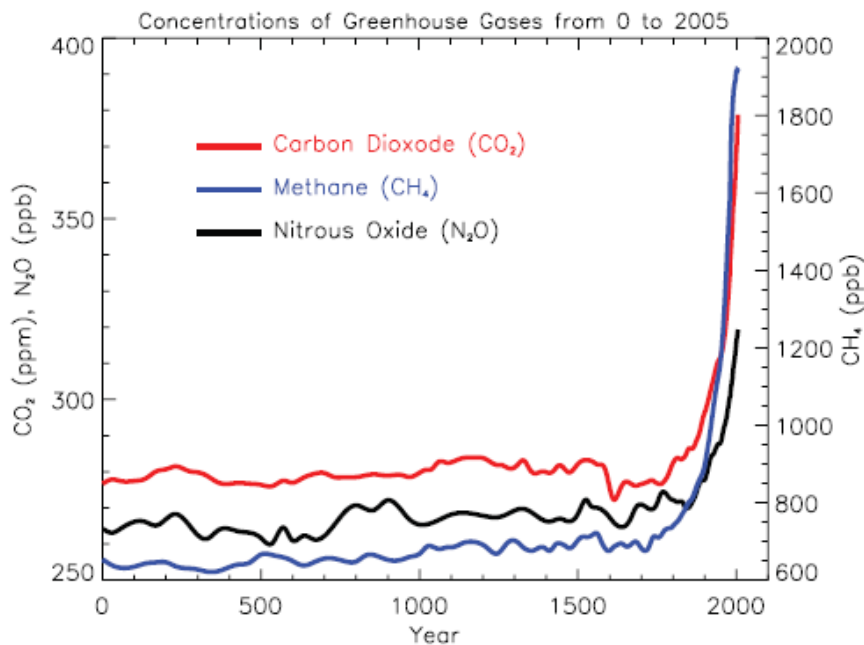


Fig. 2.1 Change in abundance of major GHGs for the last thousand years (IPCC, 2007).

According to IPCC (2007), the global average land surface air temperature in Northern and Southern Hemisphere changes for 1901 to 2000 are $0.71 \pm 0.31^{\circ}\text{C}$ and $0.52 \pm 0.13^{\circ}\text{C}$,

respectively. Because the gaseous composition of the Earth's atmosphere is changing at an accelerated rate due to the influence of human activities, which can promote global land surface air temperature increase. Results of observation showed the atmospheric concentrations of GHGs have increased from pre-industrial values to current values in 2005; about 280 to 379 ppm for CO₂, 715 to 1774 ppb for CH₄, and 270 to 319 ppb for N₂O (**Fig. 2.1**). These gases continue to increase as a result of human activities (IPCC, 2007; Solomon *et al.*, 2007).

Climate change will alter soil temperature and soil moisture both of which having strong impact on soil respiration (Raich *et al.*, 2002). Changes in soil moisture induced by climate change frequently offset or exacerbate soil temperature driven effects, particularly on soil CO₂ emissions (Saleska *et al.*, 1999). The enhanced production and reduced consumption of naturally occurring GHGs are responsible for approximately 90% of the global warming and climate change phenomenon (Solomon *et al.*, 2007). The fluxes and dissolved GHG concentrations measurement in the world as shown in **Table 2.1**. In the north peatlands area, the CO₂, CH₄, and N₂O fluxes ranged from -6.0 to 210 mgC m⁻² h⁻¹, -0.012 to 1200 mgC m⁻² h⁻¹, and -1.9 to 963 μgN m⁻² h⁻¹, respectively. Meanwhile, in north peatlands area the dissolved CO₂, CH₄, and N₂O concentrations ranged from 0.1 to 6.5 mmol L⁻¹, 0.01 to 232 mmol L⁻¹, and 3.8 to 28.2 μmol L⁻¹, respectively. Then, in the tropical peatlands area, the CO₂, CH₄, and N₂O fluxes ranged from 46 to 533 mgC m⁻² h⁻¹, -0.000453 to 8.32 mgC m⁻² h⁻¹, and -1.24 to 4298 μgN m⁻² h⁻¹. The dissolved CO₂, CH₄, and N₂O concentrations in tropical peatlands ranged from 0 to 3.290 mmol L⁻¹, 0.00001 to 1.539 mmol L⁻¹, and 0 to 3.3 μmol L⁻¹, respectively. In the south peatlands area, the CH₄ fluxes ranged from -20 to 85.6 mgC m⁻² h⁻¹ and the dissolved CH₄ concentrations ranged from <0.001 to 0.166 mmol L⁻¹.

2.2 Tropical peatlands and their characteristics

2.2.1 Distribution of tropical peatlands

Tropical peatlands is a wetland ecosystem and classified as Histosols, which has a surface layer containing more than 30% organic matter in 40 cm of the upper 80 cm of the profile (i.e. peat; FAO-UNESCO, 1990). It is a unique ecosystem, which is one of the Earth's most spatially efficient C sinks and largest long-term repositories of terrestrial organic C (Page *et al.*, 2011a). Peat C accumulates as a result of a positive net imbalance between high

tropical ecosystem primary production and incomplete organic matter decomposition in permanently saturated soil conditions (Hooijer *et al.*, 2010; Wösten *et al.*, 2008).

Although tropical peatlands are found in all humid tropical regions, the largest area and C storage is located in Southeast Asia (Page *et al.*, 2011a). According to current estimations, peatlands in this region cover an area of about 25 Mha and store 68.5 Pg C in the peat. This regional peat C store is estimated to be 77% of the C in all tropical peatlands and about 11–14% of the global peatland C pool (Page *et al.*, 2011a).

Indonesian tropical peat swamp forest is a unique wetland ecosystem and an important natural resource with considerable environmental and economic value (Rieley and Page, 1997). Indonesia has the largest area of peat swamp forest in the tropics, covering an estimated 20.7 Mha (range 16–27 Mha) (Page *et al.*, 2011a; Rieley *et al.*, 1996a and b; Sorensen, 1993) and distributed mainly across Sumatra (4.7–9.7 Mha), Kalimantan (3.1–6.3 Mha) and Irian Jaya (8.9 Mha) (Rieley *et al.*, 1996a; Silvius, 1989). In Central Kalimantan, around 3 Mha or 20% of the area is peatland (Hooijer *et al.*, 2006) and most of this remained pristine up to the beginning of the 1990s (Siegert *et al.*, 2001) (**Fig. 2.2**).

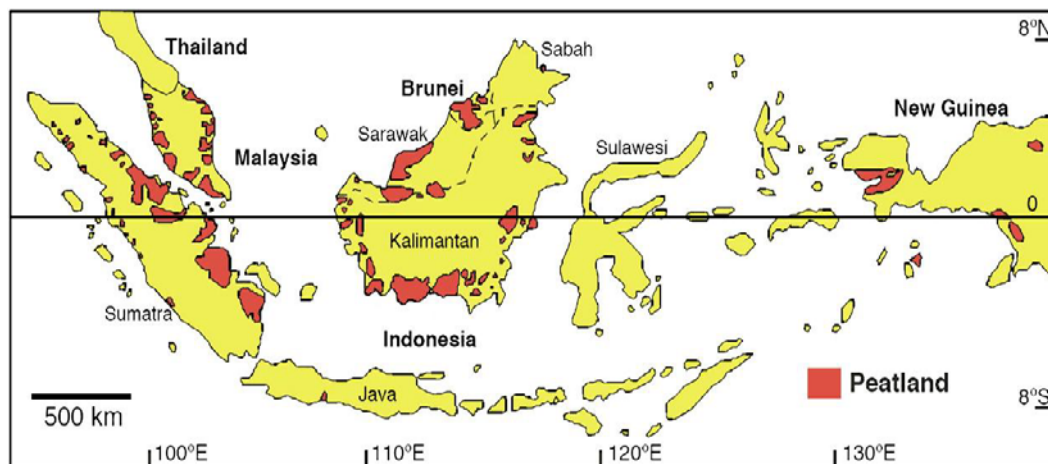


Fig. 2.2 Map of peatland distribution in the Southeast Asia (Page *et al.*, 2004).

2.2.2 Tropical peatlands under natural conditions

In a natural state, tropical peat swamp forest are characterized by dense forest vegetation, with a thickness varying from 0.3 to 20 m and a ground water table that is at or close to the ground surface throughout the year (Anderson, 1983; Hirano *et al.*, 2009; Page *et al.*, 2004; Takahashi *et al.*, 2002; Wösten *et al.*, 2008). The peat accumulation rates can be rapid, averaging 4–5 mm yr⁻¹, with some sites as fast as 5–10 mm yr⁻¹ (Maas, 1996). These rates are significantly faster than in most temperate and boreal peatlands, which accumulate

less than 1 mm yr^{-1} , often slower than 0.5 mm yr^{-1} (Gorham, 1991; Gorham *et al.*, 2003). Recent investigations (Page *et al.*, 2004) have revealed that initiation of contemporary peat deposits in Southeast Asia occurred in the Late Pleistocene ($\sim 30,000\text{--}24,000 \text{ }^{14}\text{C yrs BP}$). Accumulation was most rapid in the early Holocene ($\sim 9600\text{--}7000 \text{ }^{14}\text{C yrs BP}$, $\sim 11,000\text{--}8000 \text{ cal yrs BP}$) and continued at a reduced rate until the present day (Neuzil, 1997).

The accumulation and long-term maintenance of peat C requires a continuous supply of organic matter and a water table level that is at or close to the peatland surface throughout the year, which makes anaerobic conditions within the peat deposit (Hirano *et al.*, 2007 and 2009; Jauhiainen *et al.*, 2008 and 2005; Suzuki *et al.*, 1999). For peat accumulation to occur, the average rate of C sequestration must exceed that in decomposition losses. (Jauhiainen *et al.*, 2005, 2008, and 2010). Aboveground plant production is believed to be the primary source of peat (Clymo, 1983). According to Rieley and Ahmad-Shah (1996a and b), tropical peatlands are predominantly forested with no moss cover. Tropical peat has low bulk density ($\sim 0.1 \text{ g cm}^{-3}$) compared to mineral soils, being formed of $\sim 10\%$ tree remains and 90% water (Hooijer *et al.*, 2010), and is $50\text{--}60\%$ C by dry weight (Neuzil, 1997; Page *et al.*, 1999; 2011a). The proportion of lignin in tropical peat may be up to 75% on a dry mass basis (Hardon and Polak, 1941, as cited in Andriessse, 1988).

Aboveground biomass in tropical ombrotrophic peat swamp forest is typically in range of 100 to 250 tC ha^{-1} (Murdiyarso *et al.*, 2010; Page *et al.*, 2011a), whereas the combined above- and below-ground biomass of undrained ombrotrophic boreal *Sphagnum* dominated peatland is about 85 tC ha^{-1} (Laiho *et al.*, 2003).

Surface peat is aerobic (unsaturated) only during relatively dry periods when the water table falls below the surface (Hirano *et al.*, 2007 and 2009; Jauhiainen *et al.*, 2005 and 2010). Under natural conditions, groundwater levels within peat swamp forest follow an annual cycle that involves flooding or near surface water levels during wet season and drawdown during the dry season. In the latter case, water tables vary usually between -20 and -40 cm but do not drop more than -100 cm even during prolonged dry periods (Takahashi *et al.*, 2002 and 2003; Wösten *et al.*, 2006a and b).

Most of the peatlands in Southeast Asia is ombrotrophic system, in which peatlands have a characteristically domed, convex surface and their water and nutrient supply is derived entirely from rainfall (ombrogenous), and the organic substrate on which plants grow is nutrient poor (Andriessse, 1988). The accumulation of peat deposits over time has isolated the peatland surface from mineral-rich groundwater; hence, ombrotrophic tropical peatlands are

generally acidic and nutrient-poor, receiving all water and nutrients from precipitation (Jauhiainen *et al.*, 2010; Page *et al.*, 1999 and 2004; Wösten *et al.*, 2008).

Natural lowland tropical peatlands are dominated by trees (peat swamp forest) and are important reservoirs of biodiversity, C and water. As described in **Fig. 2.3** peat swamp forest in Sebangau Area, Central Kalimantan, Indonesia. In this area, peat swamp forests are combination of the potential high biomass productivity and created the long-term ecosystems C stores. Tropical peat swamp forests in their natural state make an important contribution to regional and global biodiversity (Andriess, 1988; Page and Rieley, 1998) and provide a vital, but under valued habitat, for rare and threatened species, especially birds, fish, mammals and reptiles (Ismail, 1999) as well as to landscape functions including water storage and supply, coastal protection, erosion prevention and flood mitigation (Rieley and Page, 2005).



Fig. 2.3 Overview of peat swamp forest canopy from the top of a tower established in Sebangau area, Central Kalimantan - Indonesia (Image: Adji F.F).

2.2.3 C storage and greenhouse gas balance in tropical peatlands

Tropical peatlands make a significant contribution to terrestrial C storage because of their considerable thickness, high C content and, most importantly, their rapid peat and C accumulation rates that have often exceeded those of boreal and temperate peatlands (Immirzi and Maltby, 1992). The tropical peatland C balance is determined largely by the net balance between C uptake in photosynthesis and C release through ecosystem respiration by: a) vegetation (autotrophic respiration and resulting in CO₂ emissions from both plant foliage and root systems) and b) the organisms involved in organic matter biological decomposition (heterotrophic respiration, involving the loss of C as CO₂ and CH₄ by organisms involved in

aerobic and anaerobic decomposition of organic matter, comprising plant litter, roots and their exudates, dead animals, fungi, bacteria and the peat it self; **Fig. 2.4**). In addition, smaller amount of C is leached out from the system in drainage runoff as dissolved organic C (DOC) or particulate organic C (POC; Moore *et al.*, 2011).

Tropical peatlands contain a large amount of organic C and nitrogen (N) under waterlogged conditions and could, therefore, be important sources of N₂O and CO₂ after drying (Bouwman, 1990; Kasimir-Klemedtsson *et al.*, 1997; Mosier *et al.*, 1998), and a source of C released as CH₄. Under certain conditions, the cycling of N makes some tropical peatlands a source of N₂O, especially if fertilizer has been added to promote agricultural or plantation productivity (Germer and Sauerborn, 2008; Jauhiainen *et al.*, 2012; Melling *et al.*, 2007; Murdiyarso *et al.*, 2010).

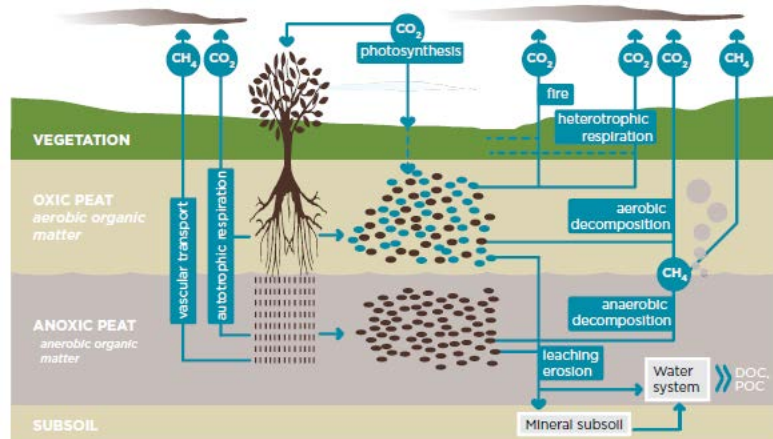


Fig. 2.4 Schematic representation of C cycle processes, flow paths, and stores in tropical peatland (after Jauhiainen *et al.*, 2001).

It is important to note that a) C cycle and GHG processes are highly dynamic and vary at all spatial and temporal scales owing to regional and local variations in macro- and micro-climate and hydrology, as well as localized variations in vegetation and peat decomposition dynamics (Hooijer *et al.*, 2012; Jauhiainen *et al.*, 2005, 2010, 2012a and b) in terms of GWP emissions, CO₂ is the most important gas emitted from drained peatlands, contributing 98% or more of the total emission of CO₂, CH₄, and N₂O (Couwenberg *et al.*, 2010; Jauhiainen *et al.*, 2012). Soil C export from forested wetlands (swamps) is dominated by soil respiration as CO₂ and CH₄ (Happell and Chanton, 1993; Krauss and Whitbeck, 2012; Yu *et al.*, 2008). As with all forests, forested wetland are long-term sinks of atmospheric CO₂ in the absence of significant disturbance (Bridgman *et al.*, 2006; Clymo *et al.*, 1998).

2.2.4 Development of tropical peatlands for agriculture

Although tropical peatlands are poorly understood scientifically, they are increasingly exploited for peat extraction, fisheries, wood harvest, and agriculture (De la Cruz, 1986; Junk, 2002; Houghton, 1999; Rieley *et al.*, 1996a). Tropical peatlands are at greater risk than remote temperate and boreal peatlands (such as in Canada and Russia), because they are often situated in densely populated low-lying coastal areas (Rieley *et al.*, 1996b). With high population growth in many tropical countries, tropical peatlands are facing increasing rates of alteration and destruction (Vijarnsorn, 1996).

Agricultural development in tropical peatlands affects the soil C pool. Thus the C balance of a tropical peatland would profoundly change. When the tree canopy is removed, C assimilation by photosynthesis is reduced (Lytle and Cronan, 1998; Nakane *et al.*, 1996). In addition, solar radiation reaching to the ground surface will increase resulting in higher diurnal temperature fluctuation in the soil. Conversion of the peat swamp to agricultural lands can lead to increases in CO₂ emissions, with cropping activities also potentially leading to increases in N₂O emissions but they can consume atmospheric CH₄ (Maljanen *et al.*, 2003a; Nykänen *et al.*, 1995; Roulet, 2000; Silvola *et al.*, 1996). However, CH₄ emissions from drainage ditches can be still significantly higher (Minkkinen *et al.*, 1997; Roulet, 2000).

Relatively large losses of organic C can also be observed when native vegetation is replaced by a lower biomass agricultural species, because this means management practices that accelerate organic C loss such as cultivation or burning (Hirano *et al.*, 2007; Howe *et al.*, 2009; Nykänen *et al.*, 1995). The important thing is that agricultural practices often replace diverse natural ecosystems with single species ecosystems. A recent research has shown that ecosystems with high plant diversity were better able to sequester C and N than ecosystems with reduced biodiversity (Lazaroff, 2001).

Deforestation and development of peatland usually accompanied the drainage of water from that areas, so much of the C has been lost by aerobic peat decomposition. In addition, the water level was likely to become uncontrollable during an exceptionally long dry period, which was frequently observed in El Niño and Southern Oscillation (ENSO) years (Hirano *et al.*, 2007 and 2012; van der Werf *et al.*, 2008), and to make the peat severely dry and susceptible to fire. Drained organic soils for agriculture have a special significance in the atmospheric N₂O load (Flessa *et al.*, 1998; Kasimir-Klemedtsson *et al.*, 1997; Maljanen *et al.*, 2003b). As drainage increases peat mineralization rates and NO₃⁻ availability, the potential

for N₂O production becomes significantly greater (Hadi *et al.*, 2000; Jauhiainen *et al.*, 2012). In addition, applications of N fertilizer generally serve to increase N₂O emission rates (Hadi *et al.*, 2000; Takakai *et al.*, 2006) although this increase can be a transient phenomenon if high N availability is not maintained (Jauhiainen *et al.*, 2012). The mechanism and influences of drainage in tropical peatland are described in detail below.

2.2.5 Impacts of peatland drainage

Because organic C sequestration in wetlands is dependent on slow decomposition rates, drainage and the associated increases in O₂ diffusion into wetland sediments often increase ecosystem respiration, leading to a net oxidation of organic material (Bridgham and Richardson, 1992; Freeman *et al.*, 1993a and b; Furukawa *et al.*, 2005; Hirano *et al.*, 2007; Jauhiainen *et al.*, 2008; Kluge *et al.*, 2008; Martikainen *et al.*, 1995). The rate at which this loss occurs is largely dependent on the depth of drainage and the subsequent land use. Drainage to greater depths increases the area of oxidation and, thus tends to increase C loss (Armentano and Menges, 1986; Furukawa *et al.*, 2005).

In general, lower water table positions with climate change are expected to increase decomposition rates by promoting aerobic microbial respiration (Freeman *et al.*, 2004). Degradation of tropical peatlands leads to release of C and a reduction in the size of their C stores (Hooijer *et al.*, 2006 and 2010; Jauhiainen *et al.*, 2005 and 2008; Page *et al.*, 2002; Rieley *et al.*, 2008). The exposure of previously saturated peat to aerobic conditions leads to the rapid transfer of historically accumulated and previously stable C to the atmosphere (Delft Hydraulics, 2006; Hooijer *et al.*, 2010 and 2012; Renewable Fuels Agency, 2010) (**Fig. 2.5**). In addition, physical and chemical disturbances such as water table level and redox state may change CH₄ and CO₂ production rates in water logged peat (Aerts and Ludwig, 1997; Hall *et al.*, 1996; Kelly and Chynoweth, 1980; Kettunen *et al.*, 1999; Öquist and Sundh, 1998). Peat CO₂ emissions are generally highest in the initial stages of drainage, owing to the rapid decomposition of a limited pool of labile C, but may decline over time as the relative amount of recalcitrant C compounds increases (Couwenberg *et al.*, 2010; Hooijer *et al.*, 2012; Verwer *et al.*, 2008).

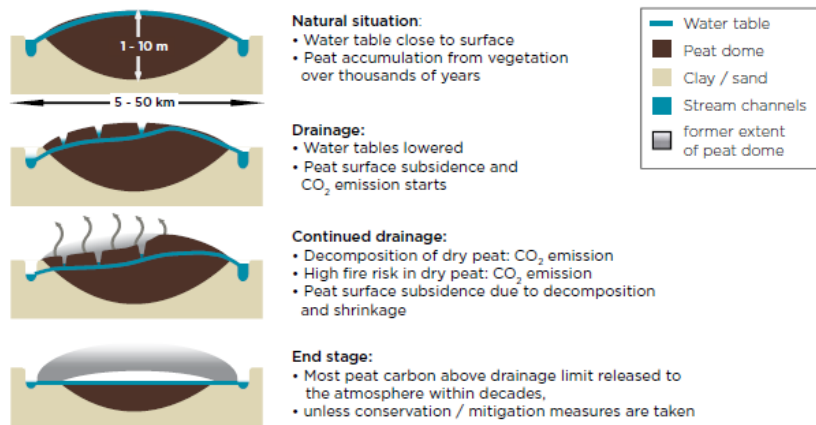


Fig. 2.5 Schematic illustration of drainage effects on a peatland dome (modified from Delft Hydraulics, 2006).

In addition, previous studies have demonstrated that the 40 cm or lower of water table level is considered to be hazardous, in where the surface peat can become sufficiently dry for ground fires to ignite and take hold within the peat (Usup *et al.*, 2004). Current C emissions from drained and fire-affected peatlands in Southeast Asia have been estimated to be of the order of -360 MtC yr^{-1} ; -170 MtC yr^{-1} from drainage-related peat decomposition (Delft Hydraulics, 2006) and 190 MtC yr^{-1} from peat fires (Page *et al.*, 2002; van der Werf *et al.*, 2008). Hirano *et al.* (2007) found a $-600 \text{ gC m}^{-2} \text{ yr}^{-1}$ ($-6 \text{ tC ha}^{-1} \text{ yr}^{-1}$) net positive C balance (peat loss) in drained peat swamp forest in Central Kalimantan, Indonesia, during the dry El Niño year of 2002, although this loss was nearly halved in wet years due to a higher water table. The C losses on this scale contribute significantly to atmospheric C loading and anthropogenic climate change processes (Page *et al.*, 2002 and 2011a and b). The long-term instability of the large amount of C stored in tropical peatlands is of major concern within the context of contemporary climate change (Raupach and Canadell, 2010).

The drainage of previously saturated peat immediately initiates the subsidence of the peatland surface and leads to a reduction in peat volume (Couwenberg *et al.*, 2010; Delft Hydraulics, 2006; Drexler *et al.*, 2009; Hooijer *et al.*, 2010; Wösten *et al.*, 1997). Subsidence is a function of the processes of peat consolidation, shrinkage, and compaction, and the decomposition (oxidation) of previously water saturated peat under aerobic conditions (Delft Hydraulics, 2006; Hooijer *et al.*, 2010 and 2012; Wösten *et al.*, 1997; Verwer *et al.*, 2008). Subsidence rates are rapid in the first one to two years following drainage, as the peat consolidates owing to increased overburden resulting from a loss of buoyancy (Hooijer *et al.*, 2012) and can result in initial subsidence rates of more than 0.5 m yr^{-1} (Hooijer *et al.*, 2012;

Wösten *et al.*, 1997). Following this primary stage of subsidence, a secondary phase of irreversible shrinkage and compaction of the peat together with rapid rates of peat decomposition leads to a slower but constant rate of subsidence (Hooijer *et al.*, 2012; Wösten *et al.*, 1997). The processes of consolidation, shrinkage, and compaction are entirely physical, and no C is lost, but peat bulk density (and C concentration) increases with time since drainage (Delft Hydraulics, 2006; Hooijer *et al.*, 2010 and 2012; Wösten *et al.*, 1997).

Lower water table positions will cause increases in soil temperature as the heat capacity of nearly saturated peat is approximately $4.02 \times 10^6 \text{ J m}^{-3} \text{ K}^{-1}$ versus that of dry peat, which is approximately $0.58 \times 10^6 \text{ J m}^{-3} \text{ K}^{-1}$ (Oke, 1987). Therefore, drier conditions could indirectly stimulate decomposition rates by affecting thermal regimes (Laine *et al.*, 2006; Lieffers, 1988). Increased nutrient availability with faster soil mineralization rates under warmer, drier soil conditions also could have consequences for vegetation.

2.3 Soil CH₄ flux

Atmospheric concentrations of CH₄ have increased by about 150% since pre-industrial time, although the rate of increase has been declining (Bartlett and Harriss, 1993; Watson *et al.*, 1990). About 70% of CH₄ production arises from anthropogenic sources and about 30% from natural sources. Currently the estimations of CH₄ potential from tropical peatland/swamps, upland soils, oceans, and lakes are 127.6, -30, 9.1, and 30 TgCH₄ yr⁻¹, respectively (EPA, 2010). The actual amount of CH₄ emitted to the atmosphere depends on the balance between CH₄ production and consumption as well as the CH₄ transport efficiency (Couwenberg, 2009). CH₄ is removed from the atmosphere by reacting with the hydroxyl radical (OH) and is ultimately converted to CO₂. Minor removal processes also include reaction with Cl⁻ in the marine boundary layer, a soil sink, and stratospheric reactions. Increasing emissions of CH₄ reduce the concentration of OH, a feedback that may increase methane's atmospheric lifetime (IPCC, 2001).

2.3.1 Magnitude and variations of soil CH₄ flux

Soils which are permanently flooded are strong sources for atmospheric CH₄ in all climate zones of the world. The magnitude of fluxes from such ecosystems is in the range of <5 to <200 mgCH₄ m⁻² d⁻¹ and can exceed 20 mgCH₄ m⁻² h⁻¹ (Delmas *et al.*, 1992; Moore *et al.*, 1990a and b; Scott *et al.*, 1999). CH₄ uptake by soils is also observed in all climate zones of the world on condition that the uppermost soil layers are well aerated. Generally, the CH₄

uptake rates by soils are mostly $<300 \mu\text{gCH}_4 \text{ m}^{-2} \text{ h}^{-1}$ and are approximately one magnitude lower than the rates of CH_4 emissions from flooded forest soils under comparable climate conditions. So far, different ecosystems on tropical peatland have not been evaluated with regard to the magnitude of CH_4 fluxes between soils and the atmosphere. As at the moment, the IPCC estimate of CH_4 emission from tropical peatland is based on the estimates of Aselmann and Crutzen (1989) and Matthews and Fung (1987) in which the tropical peatland was recognized as forest bogs. Based on the results of a study in tropical peatlands, the forest floor was a weak CH_4 sink (-0.208 to $-0.368 \text{ g m}^{-2} \text{ yr}^{-2}$; -13.0 to $-22.9 \text{ mmol CH}_4 \text{ m}^{-2} \text{ yr}^{-1}$) and the deforested site a comparable CH_4 source (0.197 to $0.275 \text{ g m}^{-2} \text{ yr}^{-1}$; 12.3 to $17.1 \text{ mmol CH}_4 \text{ m}^{-2} \text{ yr}^{-1}$) (Jauhiainen *et al.*, 2008).

CH_4 flux is highly variable at spatial scales ranging from a meter to several kilometers, and the environmental and biological controls on this variability are poorly understood. Many studies have attempted to explain the spatial variation of CH_4 emissions within wetlands in relation to changes in environmental factors, i.e. temperature, water depth, and topography (Heyer *et al.*, 2002; Moore *et al.*, 1990a and b; Waddington *et al.*, 1996). Most attempts, however, have failed to relate spatial variation in abiotic factors to that in CH_4 emissions. In recent years, biotic factors such as vegetation are considered to control CH_4 emissions from wetlands, because aquatic plants affect the production, consumption, and transport of CH_4 (Joabsson and Christensen, 2001; Whitting and Chanton, 1992). For example, in the Luanhaizi wetland, spatial variation of CH_4 flux was characterized by the dominant plants that varied in life form, density, and biomass of shoots, gas-transporting mechanism, and root-rhizome architecture (Hirota *et al.*, 2004).

The temporal variations water table level is of importance for the spatial variation, with soil CH_4 emissions generally being higher from wetter soils, and little temporal variations in CH_4 fluxes from the forest floor in drained organic soils within water table level and air temperature (Von Arnold *et al.*, 2005). The annual patterns of CH_4 fluxes are influenced by latitude, with arctic, boreal, and some temperate regions characterized by pronounced CH_4 emissions, which are governed by seasonal variability in temperature affecting water availability, production of substrate precursors and microbial activity (Whalen, 2005).

2.3.2 Soils as a source or a sink of atmospheric CH_4

Soils can function as net sinks or as net sources for atmospheric CH_4 . The exchange of CH_4 between soils and the atmosphere is the net result of simultaneous production of CH_4 in

predominantly anaerobic zones of the soil and the oxidation of CH₄ in predominantly oxic zones of the soil. Production and consumption of CH₄ within the soil profile is controlled by several environmental factors: O₂ and substrate availability, soil properties, and climate. The primary control if soils are sinks or sources for the atmospheric CH₄ is O₂ availability in the soil profile.

In waterlogged soils, CH₄ production, the final step of anaerobic matter decomposition, is primarily controlled by substrate availability and soil redox potential (Conrad, 1996; Schimel *et al.*, 1993). In nature the dominating substrates for methanogenesis are acetate and CO₂-H₂ (Knowles, 1993). The availability of these substrates and the magnitude of CH₄ production are suggested by the net primary production in an ecosystem (Aselmann and Crutzen, 1989). Further controllers of CH₄ production are soil pH (Knowles, 1993) and temperature (Conrad, 1996).

Since CH₄ is mainly produced in a certain depth in wetland soils, it must be transported to the soil surface prior to its emission into the atmosphere. During this process CH₄ can be oxidized by methanotrophic bacteria. Several studies in wetland ecosystems have shown that 40 to 95% of total CH₄ produced in the soil (Butterbach-Bahl *et al.*, 1997; Galchenko *et al.*, 1989) is oxidized either in the surface aerobic soil layer, or in oxygenized zones around plant roots (Gilbert and Frenzel, 1998), before it is released to the atmosphere. Diffusion of CH₄ through water saturated sediments have been shown to be of a minor importance as compared to the other emission pathways (Butterbach-Bahl *et al.*, 1997), since diffusion of gases in water is approximately a factor of 10,000 lower than in air. For wetland ecosystem, it has been shown that plant mediated transport is the main mechanism by which CH₄ is emitted from the soil to the atmosphere (Butterbach-Bahl *et al.*, 1997; Schimel *et al.*, 1993; Whiting and Chanton, 1993).

Well aerated soils are significant sinks for atmospheric CH₄ (Adamsen and King, 1993; Castro *et al.*, 1995; Crill, 1991; Keller *et al.*, 1983; Yavitt *et al.*, 1990). Main uptake activity of atmospheric CH₄ has been shown to be concentrated in the uppermost mineral horizon of soils (Adamsen and King, 1993; Bender and Conrad, 1994; Brumme and Borcken, 1999; Saari *et al.*, 1998; Steinkamp *et al.*, 2001), whereas the organic layer has low CH₄ oxidation activity or may even show low rates of net-production of CH₄ (Adamsen and King, 1993; Sexstone and Mains, 1990). The reasons for this wide spread observations that the main CH₄ oxidation activity of soils is located in the uppermost mineral soil layer and not in the organic layer which is directly exposed to the atmosphere are still uncertain.

Atmospheric CH₄-oxidation has been demonstrated to be extremely sensitive to increased inorganic-N concentrations in the soil (Castro *et al.*, 1995; Macdonald *et al.*, 1997; Sitaula *et al.*, 1995; Steudler *et al.*, 1989). Some researchers had assumed that higher NH₄⁺ content in the organic layer as compared to the uppermost mineral layer inhibit CH₄ oxidation (Bender and Conrad, 1994; Conrad, 1996; Schnell and King, 1994). Laboratory experiments had indicate that the inhibitory effect of increased inorganic-N contents on CH₄ activity may be due to a competitive inhibition of the enzyme CH₄ monooxygenase by NH₃⁺ (Adamsen and King, 1993; Dunfield and Knowles, 1995; King and Schnell, 1994) or the production of toxic products (NO₂⁻ and NH₂OH) produced during enhanced NH₄⁺ oxidation (King and Schnell, 1994). Furthermore, the organic layer may contain further compounds that may control CH₄ oxidation activity. As shown by Amaral and Knowles (1998), CH₄ oxidation rates can be inhibited by increasing concentrations of monoterpene. They had suggested that soil monoterpene distribution may be responsible for stratification of CH₄ oxidation activity in soils.

2.3.3 Environmental factors controlling CH₄ flux

Empirical models have been tried to explain the variation in CH₄ exchange rates and to reflect average differences in environmental and ecological variables within and among peatlands (Moore *et al.*, 1998). At smaller scales, much less of the variation of CH₄ exchange rates can be explained by environmental and ecological variables (e.g. Bellisario *et al.*, 1999; Kettunen *et al.*, 1996; Moore, 1994; Shannon and White, 1994). Changes of temperature (Benstead and Lloyd, 1996; Speppard *et al.*, 2007), water table position (Benstead and Lloyd, 1994; Daulat and Clymo, 1998), and season (Benstead and Lloyd, 1996) are the causes of variability in the concentration and distribution of CH₄ following in situ investigation.

2.3.3.1 Land use

CH₄ uptake rates of soils are strongly influenced by human activities. This is strongly supported by the results of N-fertilization experiments (Castro *et al.*, 1995; Gullledge *et al.*, 1997; Sitaula *et al.*, 1995; Steudler *et al.*, 1989; Whalen and Reeburg, 2000) whereby the rates of atmospheric CH₄ oxidation is strongly reduced by increasing N availability. CH₄ oxidation in soils has been shown to be extremely sensitive to physical disturbance of the soil structure as may occur if land use is changed from intact forest to agricultural land. This land

conversion has been shown to reduce CH₄ uptake rates in temperate regions by at least 60% (Dobbie *et al.*, 1996; Priemé *et al.*, 1997), whereas for soils of the wet tropics such as conversion of forest to pastures may even turn soils from sinks to sources of atmospheric CH₄ (Keller and Reiner, 1994). In contrast to the short-term effects of land use change from forestry to agriculture on CH₄ fluxes, it may take decades after conversion of arable land to forest to regain CH₄ uptake rates which are comparable to undisturbed forest sites (Keller and Reiners, 1994; Priemé *et al.*, 1997). The reason for these discrepancies is still poorly understood, but may be related to major changes in the physical structure of soils, e.g. compaction of the soil (Dobbie *et al.*, 1996; Ojima *et al.*, 1993).

Vascular plants can enhance CH₄ emission rates by serving as a conduit for gas by means of aerenchyma system by which CH₄ produced in anaerobic environments bypasses the oxidation zone where less than 10% of CH₄ may diffuse across the oxic line (unsaturated zone) (Frenzel *et al.*, 1990). Thereby, plant-mediated transport is known to be of the most importance, accounting for up to 90% of total CH₄ emission (Banker *et al.*, 1995; Holzapfel-Pschorn *et al.*, 1986; Jia *et al.*, 2001; Yagi *et al.*, 1996). Besides transporting CH₄, plants can also stimulate CH₄ production by providing substrates for methanogens and facilitate CH₄ oxidation by delivering O₂ downward into the rhizosphere for methanotrophs (Jia *et al.*, 2001; van der Nat and Middelberg, 1998). Moreover, plant species also greatly affected the magnitude of CH₄ emissions from wetlands (Crill *et al.*, 1988; Whalen and Reeburgh, 1988) especially when plant species varied from bryophytes, which lack vascular structure, to vascular plants (Bubier, 1995). Verville *et al.* (1998) found that, on a local scale, vegetation composition had a greater effect on CH₄ emissions than direct manipulation of air and soil temperature.

Up to date, the majority of tree species that possess adaptive structures to facilitate O₂ ingress are also capable of mediating CH₄ egress. The previous studies have shown six of species tree: Elaeocarpaceae (*Elaeocarpus mastersii*), Ebenaceae (*Diospyros bantamensis*), Myrtaceae (*Tristaniaopsis* sp. 2), Clusiaceae or Guttiferae (*Mesua* sp. 1), Lauraceae (*Litsea elliptica*), Annonaceae (*Xylopius fusca*) emitted CH₄ from their stems at the Borneo Island (Pangala *et al.*, 2013). In addition, Pangala *et al.* (2013) also reported multiple regression analysis indicates that stem diameter, wood specific density and concentration of CH₄ dissolved in pore water explain up to 80% of stem CH₄ flux variations.

2.3.3.2 Drainage

Water table depth and redox potential exert strong controls over CH₄ emissions from wetlands (Funk *et al.*, 1994; Moore and Dalva 1993; Wang *et al.*, 1993). CH₄ emission is the balance of two counteracting processes: methanogenesis in anoxic conditions and the oxidation of the generated CH₄ (Bastviken *et al.*, 2002; Minkkinen and Laine, 2006). The effects of water table drawdown on CH₄ emissions are more complex as there may be enhanced CH₄ production through root exudates and litter associated with denser vascular plant colonization, but there may also be enhanced transport of CH₄ directly to the atmosphere (Strack *et al.*, 2008). However, while CH₄ production may be enhanced, CH₄ oxidation may also be enhanced due to the transport of O₂ to the rooting zone (Sundh *et al.* 1995; Strack *et al.*, 2008). Water table exerts a major control on CH₄ flux from wetlands by creating aerobic and anaerobic zones in the soil profile, and thus the potential for CH₄ consumption and production (Moore and Roulet, 1995).

Lowering of water table in peatlands has occurred over a large area in temperate, boreal and subarctic regions (Armentano and Menges, 1986; Gorham, 1991). It would be expected that this would increase the thickness of the zone of CH₄ consumption and decrease the thickness of the zone of CH₄ production in the peat profile, resulting in a reduction in CH₄ emission rates. Thus drained peatlands have very small CH₄ consumption rates (generally <2 mgCH₄ m⁻² d⁻¹), similar to the rates of CH₄ consumption observed in forest, grassland and upland tundra soils (Adamsen and King, 1993; Steudler *et al.*, 1996; Striegl *et al.*, 1992; Whalen and Reeburg, 1990; Whalen *et al.*, 1991).

Under saturated condition, production of CH₄ in tropical peatland was increased due to higher water table level. In these conditions O₂ concentrations was low corresponding to anaerobic processes. Previous studies have demonstrated that net CH₄ emissions from wetlands are largely controlled by plant-mediated transport (Kelker and Chanton, 1997). Although significant quantities of CH₄ are produced in the peat, CH₄ is typically not released at high rates from the peat surface to the atmosphere because methanotrophic bacteria oxidize CH₄ at the oxic-anoxic interface in surface soil and within the rhizosphere (Couwenberg *et al.*, 2010). The role of trees in the CH₄ cycle should not, however, excuse deforestation, because the measurement of tree-mediated CH₄ flux expressed in CO₂ equivalents represents <2% of total C emissions from deforested tropical peat forest (Hirano *et al.*, 2007).

2.3.3.3 Other environmental factors

As for every biological process, microbial CH₄ oxidation will increase with increasing temperatures. Several studies have shown that soils under <10°C are subject to temperature change (Castro *et al.*, 1995; Crill, 1991; Steinkamp *et al.*, 2001). At temperatures >10°C, other factors such as soil moisture have been observed to be more important for the magnitude of CH₄ uptake than soil temperature (Castro *et al.*, 1995; Steinkamp *et al.*, 2001).

Atmospheric CH₄ uptake is largely controlled by gas diffusion resistance within the soil. Therefore, the structure of the organic layer (Brumme and Borken, 1999; Dong *et al.*, 1998) and the texture of the mineral soil (Boeckx *et al.*, 1997; Dorr *et al.*, 1993) have a huge impact on the magnitude of rates of CH₄ oxidation in soils.

Moisture availability appears to act as a switch turning CH₄ emissions on and off. The presence of a large saturated zone creates a high potential for methanogenesis. Soil moisture directly influences the gas permeability of soils and therefore also will be affected the exchange of gases between soils and the atmosphere. Thus, an increase in soil moisture will reduce CH₄ oxidation activity by limiting substrate availability (Brumme and Borken, 1999; Castro *et al.*, 1995; Gullledge and Schimel, 1998). Furthermore, high soil moisture contents will also decrease O₂ availability in the soil profile, thereby inhibiting the process of CH₄ oxidation, since for the oxidation of CH₄ to methanol by the enzyme CH₄ monooxygenase, O₂ is a prerequisite (Gullledge and Schimel, 1998; Knowles, 1993).

2.4 Soil CO₂ flux

The atmospheric CO₂ is currently increasing at 0.4% per year since 1980 and now constitutes approximately 367 ppm compared to 280 ppm in pre-industrial times in 1750, a 31% increase (IPCC, 2001; Watson *et al.*, 1992). The rate of increase over the past century is unprecedented at least during the past 20,000 years. The IPCC had definitively states that the present atmospheric CO₂ increase is caused by anthropogenic emissions of CO₂ (IPCC, 2001).

The terrestrial ecosystems are currently acting as a global sink for C. Forest ecosystems sequester larger amount of C compared with croplands (Houghton *et al.*, 2000; Pete Smith, 2004; Valentini *et al.*, 2000). In its second assessment, IPCC (1996) also stated that the increased amount of CO₂ in the atmosphere is leading to climate change and will produce, on average, a global warming of the Earth's surface because of its enhanced greenhouse effect, although the magnitude and significance of the effects are not fully resolved.

Soils are an important component of the global C budgets, containing about twice the amount of C as the atmosphere, i.e. 1500 Pg of organic C in soils versus 750 Pg of C in the atmosphere (Eswaran *et al.*, 1993). Thus even a minor fluctuation in soil C pool will lead to a great variation in global pool and managing soil C sequestration to mitigate climate change (Houghton *et al.*, 1999; Lal *et al.*, 2004; Schlesinger, 1999).

2.4.1 Magnitude and variations of soil CO₂ flux

Despite the large number of studies, there is still a large uncertainty about the magnitude of soil CO₂ flux and its contribution to the global CO₂ budgets and the factors controlling it. In order to estimate soil CO₂ flux more accurately, more understanding is needed on the processes controlling soil CO₂ flux. There are many published reports of soil CO₂ flux from tropical peatlands (e.g. Chimner, 2004; Hirano *et al.*, 2009; Inubushi *et al.*, 2003; Jauhiainen *et al.*, 2001 and 2008). According to the previous studies, Chimner (2004) suggested that the soil respiration rates were higher in the warmer Micronesian peatlands (2.15–2.54 $\mu\text{mol m}^{-2} \text{s}^{-1}$) than in the cooler Hawaiian montane peatlands (0.83–1.81 $\mu\text{mol m}^{-2} \text{s}^{-1}$), resulting in higher annual CO₂ flux rates compared to other non tropical peatlands. Then, converting secondary forest peatland to paddy field tended to increase annual fluxes of CO₂ to the atmosphere from 1.2 to 1.5 $\text{kgCO}_2\text{-C m}^{-2} \text{yr}^{-1}$, while changing land use from secondary forest to upland (abandoned agricultural land) tended to decrease these gas emissions from 1.2 to 1.0 $\text{kgCO}_2\text{-C m}^{-2} \text{yr}^{-1}$ (Inubushi *et al.*, 2003).

Soil CO₂ flux exhibits a high spatial and temporal variability, which has been related to either root biomass, microbial biomass, litter amount, soil organic C, soil N, cation exchange capacity, soil bulk density, soil porosity, soil pH, or site topography (Fang *et al.*, 1998; Hanson *et al.*, 1993; La Scala *et al.*, 2000; Risch and Frank, 2006). Seasonal variations of soil CO₂ flux have often been associated with either changes in soil temperature (Anderson, 1973; Edwards, 1975; Ewel *et al.*, 1987; Fang *et al.*, 1998; Longdoz *et al.*, 2000) or changes in both soil temperature and soil water content (Davidson *et al.*, 1998; Epron *et al.*, 1999; Garret and Cox, 1973; Hanson *et al.*, 1993; Qi and Xu, 2001; Xu and Qi, 2001).

In tropical peatlands, CO₂ fluxes at the forest soil were clearly higher in comparison to the deforested area. Cumulative forest floor CO₂ fluxes (7305 to 7444 $\text{g m}^{-2} \text{yr}^{-1}$; 166.0 to 1.69.2 $\text{molCO}_2 \text{ m}^{-2} \text{yr}^{-1}$) and deforested site CO₂ fluxes (2781 to 2608 $\text{g m}^{-2}\text{yr}^{-1}$; 63.2 to 59.3 $\text{molCO}_2 \text{ m}^{-2} \text{yr}^{-1}$) did not reflect the differing hydrological conditions in the years before and after restoration of drainage (Jauhiainen *et al.*, 2008). Hirano *et al.* (2009) reported the mean cumulative hummock CO₂ fluxes from tropical peatlands were $109.0 \pm 9.0 \text{ mol m}^{-2}$

yr⁻¹ (4794±422 gCO₂ m⁻² yr⁻¹), and the soil CO₂ flux increased exponentially with soil temperature (T_s) even within an-amplitude of 4–5°C.

2.4.2 Soil CO₂ flux and the ecosystems' C balance

Soil CO₂ flux is one of the largest fluxes in the global C cycle and is a key component of ecosystem C balances (Houghton, 1995; Raich and Schlesinger, 1992; Schlesinger and Andrews, 2000). Ecosystems contribute CO₂ to the atmosphere through respiration and consume CO₂ from the atmosphere via photosynthesis. CO₂ is assimilated in photosynthesis by trees and ground vegetation and translocation to soils through several pathways (**Fig. 2.6**). Significant amounts of C are allocated to the root systems for root growth and root maintenance. When the roots die, the C is added to the soil as dead organic matter. Carbon is also added to the ground surface and humus from aboveground biomass through litter fall and leaching of dissolved organic matter from the canopy (Edwards and Harris, 1977; Kalbitz *et al.*, 2000) and from roots (Högberg *et al.*, 2001). Carbon is released from the soil to the atmosphere through the decomposition of dead organic matter and through the respiration by roots, root mycorrhizal fungi and other soil micro-organisms (Chapin III and Ruess, 2001; Gaudinski *et al.*, 2000). Some of the C is also leached out of the ecosystem dissolved in ground water especially in peatlands (Urban *et al.*, 1989; Sallantaus, 1992).

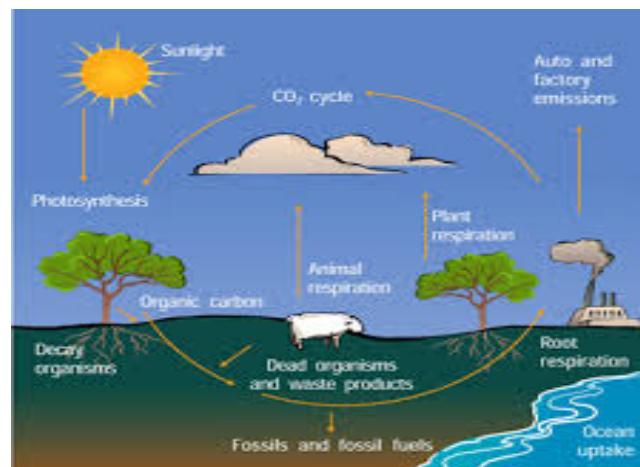


Fig. 2.6 Schemes of C fluxes and factors controlling them in the ecosystems.

The relationship between production and decomposition determines whether an ecosystem is a sink or a source of atmospheric CO₂. In a climax ecosystem, soil respiration is taken to be approximately balanced by photosynthetic uptake of CO₂ to produce biomass.

These two fluxes are of similar magnitude and changes in climate and the length of growing season can shift a forest from being a sink to be a source of C (Valentini *et al.*, 2000). Soil CO₂ fluxes are produced within the soil by heterotrophic microbial respirations and by autotrophic root respiration. To date, C emissions from soil to the atmosphere via decomposition of organic matter plus root respiration are approximately 10-fold greater than fossil fuel and deforestation sources combined (Schimel *et al.*, 2000); hence, even small changes in total emissions will affect atmospheric chemistry and heat balance.

2.4.3 Environmental factors controlling soil CO₂ flux

The conversion of forests to agriculture is responsible for a substantial increase in atmospheric CO₂ concentration (Houghton *et al.*, 1983; Houghton, 1995). Air and soil temperature, water table level, and the quality of organic substrates are the main local controls of CO₂ emissions from peatlands (Bridgham *et al.*, 1995). In laboratory studies and the theory (Clymo, 1983; Doran *et al.*, 1990; Linn and Doran, 1984), water table levels can have important effects on CO₂ emissions from peatlands, because saturated soils limit the diffusion of atmospheric O₂ into the peat, limiting microbial activity and decomposition rates. Conversely, a water table decline increases O₂ diffusion into soils allowing aerobic decomposition, which increases CO₂ emissions (Bubier, 1995; Moore and Knowles, 1989; Nykänen *et al.*, 1998; Silvola *et al.*, 1996). The previous studies have demonstrated the respiration by roots plus oxidation of rhizosphere C from mixed temperate forest can produce a large portion of total CO₂ efflux due to sensitivity of soil respiration in elevated temperatures (Boone *et al.*, 1998). While liming activities increased the CO₂ emission to 4.1 tC ha⁻¹ yr⁻¹ from a temperate forest and drastically reduced the N₂O emission to 1.5 kgN ha⁻¹ yr⁻¹ (Brumme and Beese, 1992).

2.4.4 Autotrophic and heterotrophic respirations in soils

Soil CO₂ emission, as the result of soil respiration generates mainly from autotrophic (root) and heterotrophic (microbial) activity (Janssens *et al.*, 2001). Soil microorganism release CO₂ by oxidizing organic debris and return the C assimilated by the plants back to the atmosphere. Major factors affecting microbial respiration are the amount and quality of organic C in the soil, soil temperature and soil moisture (Buyanovsky *et al.*, 1986; Davidson *et al.*, 1998; Kiese and Butterbach-Bahl, 2002; Kirschbaum, 1995; Mosier, 1998; Prescott *et al.*, 2000). These factors are highly variable, depending on the geographical location of the site, the physical and chemical properties of the soil, and type of ecosystem.

Root and rhizosphere respirations are another important component of soil CO₂ flux. Estimates on the contribution of root and rhizosphere respirations are highly variable, ranging from 10 to 90% of total flux (Bowden *et al.*, 1993; Ewel *et al.*, 1987; Hanson *et al.*, 2000; Maier and Kress, 2000; Nakane *et al.*, 1983 and 1996). Direct measurements of root and rhizosphere respirations are difficult because the measurements themselves usually affect the respirations, e.g. injuring the roots. Moreover, instantaneous measurements of root respiration are difficult to scale up to the ecosystem level because of large spatial variation in root distribution (Buchmann, 2000). The amount of root and rhizosphere respiration is dominated by the root biomass of a specific soil layer. Pietikäinen *et al.* (1999) and Widén and Majdi (2001) found the highest respiratory activities in boreal forest in organic layer close to the soil surface, where the amount of fine root biomass was also the highest. However, the rate of CO₂ production by roots at different depths depends also on the proportion of new and old roots. As the root tissue mature, there is gradual decline in respiration (Singh and Gupta, 1977). The photosynthetic activity of leaves influences the rate of root and rhizosphere respirations (Högberg *et al.*, 2001; Singh and Gupta, 1977).

2.5 Soil N₂O flux

N₂O is one of the most important anthropogenically enhanced GHGs, behind CO₂ and CH₄. It contributes about 5% to global warming (IPCC, 2001) and involved in the destruction of stratospheric ozone (Crutzen, 1970). About 70% of the total globally emitted N₂O is derived from soils (Bouwman, 1990; Conrad, 1996) and agriculture as a whole (i.e., animal excreta, denitrification of leached nitrate (NO₃⁻), etc.) contributes about 81% of the anthropogenic N₂O emissions (Brown *et al.*, 2001). Direct and indirect emissions from agricultural systems are now thought to contribute approximately 6.2 TgN₂O-N yr⁻¹ to a total global source strength of 17.7 TgN₂O-N yr⁻¹ (Kroeze and Bouwman, 1999), and tropical soils are thought to account for 20–50% of all global sources of atmospheric N₂O (Davidson and Kinglerlee, 1997).

The atmospheric concentration of N₂O has increased by 16% since 1750, from a pre-industrial value of about 270 to 314 ppb in 1998, a concentration that has not been exceeded during the last thousand years. The annual global flux to the atmosphere is estimated to be about 4 million metric tons of N₂O (Mosier *et al.*, 1998). This is equivalent to a 0.2 to 0.3 percent annual increase. This would create an adverse impact on global climate in the 21st century (IPCC, 1995; Lashof and Ahuja, 1990). It has been estimated that doubling the

concentration of N₂O in the atmosphere would result in a 10% decrease in the ozone layer and this would increase the ultraviolet radiation reaching the Earth by 20% (Crutzen and Ehhalt, 1977).

Soils are usually net sources for atmospheric N₂O, but they can also act as a sink, at least temporary (Butterbach-Bahl *et al.*, 1998; Duxbury and Mosier, 1993; Firestone, 1982; Freney *et al.*, 1978; Gasche and Papen, 1999; Papen and Butterbach-Bahl, 1999; Ryden, 1981; Slemr and Seiler, 1984). The N₂O flux is the result of dynamic production and consumption processes in soil. However, most of the N₂O produced within the soil column may be consumed and never reach the atmosphere (Arah *et al.*, 1991; Seiler and Conrad, 1981). Consumption of N₂O is due to the reduction to N₂. Oxidation of N₂O by soil catalase and peroxidase has only been in two publications, but was never investigated again (Knowles, 1985).

There have been few studies on the emission of N₂O from peatlands but pristine peatlands seem to be neither a significant source nor sink (Martikainen *et al.*, 1995; Schiller and Hastie, 1994; Roulet, 2000). To date, there are few reports of N₂O emissions from the agriculture field at tropical agricultural systems (e.g. Takakai *et al.*, 2006; Toma *et al.*, 2011), despite these systems being the predominant land use in much of the humid tropics especially for tropical peatland. But the enhancement of the N mineralization in drained peat leads to greater emissions of N₂O from nitrification and denitrification.

2.5.1 Magnitude of soil N₂O flux under different climate

The total flux of N₂O into the atmosphere from all sources is currently estimated at 18.8 Tg yr⁻¹ as N, which represents an increase since the pre-industrial era (Solomon *et al.*, 2007). This increase primarily reflects human activities, particularly agriculture and associated land use change. The IPCC estimates that about 60% of all N₂O emissions come from natural sources, but individual source estimates remain subject to significant uncertainties (Forster *et al.*, 2007). According to EPA (2010), N₂O fluxes from natural sources to the atmosphere is 12.1 TgN yr⁻¹, or 64% of the total of all emissions worldwide. The estimations of N₂O potential from natural, ocean, tropical soils, and temperate soils (forest), and grassland are 3.0 (1.0–5.0), 4.0, 3.0 (2.2–3.7), 1.0 (0.1–2.0), and 1.0 (0.5–2.0) TgN₂O yr⁻¹, respectively (Mosier *et al.*, 1998). Currently, Asia is a hotspot of N mobilization and distribution (Galloway, 2000). In Asia, the use of synthetic N fertilizers to agricultural fields has increased from 2.1 in 1961 to 40.2 TgN in 1994 and the corresponding increase of N₂O was from about 0.8 to about 2.1 TgN₂O-N, respectively (Mosier and Zhaoliang, 2000).

There is general agreement that tropical forest soils are very important N₂O sources on a global scale (Garcia-Montiel *et al.*, 2001; Keller *et al.*, 1993; Melillo *et al.*, 2001; Verchot *et al.*, 1999) and probably the most important natural source of this gas to the atmosphere. However, research work on N₂O emission from organic soil is mostly done in boreal and temperate regions. According to the current IPCC Guidelines, N₂O emissions from organic soils in cool and temperate climates are estimated to be 8 kgN₂O-N ha⁻¹. Mineralization rates of N in tropical climates are just assumed to be about 2 times greater than in temperate climates, so the current estimate from organic soil in the tropics is 16 kgN₂O-N ha⁻¹ (Penman *et al.*, 2000).

The increasing use of inorganic N fertilizers in Indonesia having a warm, tropical climate with frequent rainfall events may contribute to N₂O production significantly. Previous studies have explained that the annual N₂O emissions from the croplands in tropical peatland were significantly higher (21±5.4 to 131±59 kgN₂O-N ha⁻¹ yr⁻¹ in 2002–2003; 52±8.2 to 259±44 kgN₂O-N ha⁻¹ yr⁻¹ in 2003–2004) compared to the natural forest (0.62±0.11 kgN₂O-N ha⁻¹ yr⁻¹ in 2002–2003; 4.4±1.2 kgN₂O-N ha⁻¹ yr⁻¹ in 2003–2004), regenerated forest (0.40±0.32 kgN₂O-N ha⁻¹ yr⁻¹ in 2002–2003; 4.0±1.9 kgN₂O-N ha⁻¹ yr⁻¹ in 2003–2004), and burned forest (0.97±0.65 kgN₂O-N ha⁻¹ yr⁻¹ in 2002–2003; 1.5±0.7 kg N₂O-N ha⁻¹ yr⁻¹ in 2003–2004) (Takakai *et al.*, 2006). In addition, the N₂O emissions in conventionally cultivated croplands and unplanted/unfertilized treatments (bare soil) in each cropland at the tropical peatland varied from 10.9 to 698 kgN ha⁻¹ yr⁻¹ and 6.55 to 858 kgN ha⁻¹ yr⁻¹, respectively (Toma *et al.*, 2011).

2.5.2 Temporal and spatial variation in soil N₂O flux

Most studies in temperate soils had reported daily variations of N₂O flux rates (Bowden *et al.*, 1991 and 1993; Brumme and Beese, 1992; Lemke *et al.*, 1998; Ryden *et al.*, 1978; Velthof *et al.*, 1996). Many forest ecosystems have been demonstrated to exhibit pronounced seasonal as well as inter annual variations of N₂O (Gasche and Papen, 1999; MacDonald *et al.*, 1997; Papen and Butterbach-Bahl, 1999). The seasonal variations are mainly controlled by seasonal changes in soil water and temperature (Skiba *et al.*, 1999). Most studies revealed that the fluxes were highest during spring and summer and lowest during autumn and winter (Brumme and Beese, 1992; Butterbach-Bahl *et al.*, 2002; Gasche and Papen, 1999; Papen and Butterbach-Bahl, 1999). However, Bowden *et al.* (1993) and Brumme *et al.* (1999) had

shown that temperate forest ecosystems did not exhibit any seasonal pattern of N₂O emissions at all. The reason for these conflicting results remains to be clarified.

With respect to tropical forest ecosystems it was generally found that N₂O emission rates during the wet season were markedly higher as compared to the dry season (Breuer *et al.*, 2000; Davidson *et al.*, 1993; Garcia Méndez *et al.*, 1991; Keller and Reiners, 1994; Steudler *et al.*, 1991; Verchot *et al.*, 1999; Vitousek *et al.*, 1989). However, the highest N₂O emission rates were observed during the transition period from the wet season to the dry season at tropical rain forest in Australia (Breuer *et al.*, 2000). The weather affects both the microbial activities in the soil and gas diffusion through the soil leading to great temporal variation of the N₂O emissions (Dobbie *et al.*, 1999; Flessa *et al.*, 1998).

N₂O flux has been studied on the level of net fluxes between soil and the atmosphere in the field and in laboratory incubations of intact soil columns or soil samples from various soil horizons, and on the level of oxidative versus reductive production processes. The observed fluxes are typically log-normally distributed and cover about 4 orders of magnitude (Conrad, 2002). This may result from a highly dynamic regulation on the microbial process level causing a highly variable flux both temporally and spatially. A very distinct hot-spot phenomenon, macroscopic flux, is sometimes caused by only a small fraction of the study areas (Ambus and Christensen, 1994; Parkin, 1987). Spatial variability of N₂O emission rates varied between 94–196% (Vitousek *et al.*, 1989), 100–118% (Verchot *et al.*, 1999), and 14–132% (Breuer *et al.*, 2000).

Chamber measurements used in most studies for the determination of N₂O fluxes from soils have revealed pronounced variation of fluxes in space and time due to the interaction of environmental factors: e.g. soil temperature, WFPS, NO₃⁻, and NH₄⁺ (Gasche and Papen, 1999; Groffman *et al.*, 2000; Mosier, 1989; Papen and Butterbach-Bahl, 1999; Sitaula and Bakken, 1993; Velthof *et al.*, 1996). In studies on the landscape scale, it was reported that the topography also exerts spatial variability on N₂O fluxes because it has a strong influence on the hydrological and pedological processes that directly control denitrification (Pennock *et al.*, 1992; Van Kessel *et al.*, 1993). Topographical position, differences in WFPS and C/N ratio were found to be important controllers of N₂O emission from tropical rain forest soils (Breuer *et al.*, 2000). The spatial variability of N₂O could also be regulated by the microorganisms through its growth dynamics or enzyme production rates, when the other input variables are constant (Muller *et al.*, 1997). The effect of a single soil variable cannot be separated from other variables because the soil variables would influence each other at the

same time. Thus, the N₂O emissions show a large spatio-temporal variability (Dobbie *et al.*, 1996; Granli and Bockman, 1994; Velthof *et al.*, 1996) leading to great uncertainties in the global N₂O budget.

2.5.3 Environmental factors controlling N₂O flux

There are several environmental factors that influence the N₂O emission from agroecosystems (Mosier *et al.*, 1996). The processes that may be involved in determining the flux between soil and atmosphere of N₂O have been reviewed by several authors (Beauchamp, 1997; Bender and Conrad, 1994; Conrad, 1995 and 1996; Davidson and Verchot, 2000; Firestone and Davidson, 1989; Granli and Boeckman, 1994; Knowles, 1985; Sahrawat and Keeney, 1986). There is an agreement that chemical reactions in soil play no role in the turnover of N₂O (Bremner *et al.*, 1980). Under most soil conditions, however, the turnover of N₂O is caused by biochemical reactions taking place within microbial cells. Thus it is obvious that microbial processes ultimately cause N₂O flux between the soil and the atmosphere.

Previous studies have been described about the regulation of N₂O turnover in soil. Generally, continuous labile N availability from vegetation or labile organic compounds, water table level, soil temperature, soil moisture, and soil O₂ concentrations are the universal controlling factors for N₂O emissions, but these factors operate in different combinations and order of importance both temporally and spatially (Burgin and Groffman, 2012; Jauhiainen *et al.*, 2012a; Jenkinson, 2001; Mosier, 2001; Skiba and Smith, 2000).

2.5.3.1 Abiotic factors

In numerous studies, soil moisture has been found to be one of the most important regulators (Davidson, 1991; Luo *et al.*, 1998). It has been suggested that N₂O is the dominant product above field capacity and mainly produced by denitrification. Soil moisture affects not only the metabolism of N but also of C, which in itself is another soil variable that regulates the turnover of N₂O. Precipitation events cause an increase in soil water content, which stimulates respiratory processes and impedes gaseous diffusion in soil. This results in local consumption of O₂ thus enhancing anaerobic microbial metabolism and possibly stimulating N₂O production. On the other hand, however, diffusion of N₂O is also impeded and thus consumption of this gas is stimulated.

NO₃⁻ can also accumulate in dry soil (Davidson *et al.*, 1991) as mineralization occurs and be released with readily available C (Davidson *et al.*, 1987; Mummey *et al.*, 1994).

Subsequent precipitation may then stimulate denitrification. However, there can also be a piston effect (Definition: piston effect is described as tunnel in soil, when the tunnel (porosity) in soil not filled with the water, the air trapped in soil will be pushed out to the atmosphere by diffuse process), with rainwater pushing out N₂O trapped in the soil (Machefert *et al.*, 2004). Some researchers have reported enormous pulses of N₂O emission from tropical forest soils after rewetting of the soil at the transition period between dry season and wet season with the onset of rainfall indicating that lack of water limited microbial production of N₂O (Davidson *et al.*, 1991 and 1993; Serca *et al.*, 1998).

Soil temperature influences the physiological activity of soil microorganism and it can alter microbial activity in soil, rates of organic decomposition and ecosystem C storage (Zak *et al.*, 1999). The enhanced microbial respiration and N mineralization are related to larger substrate pools and the increases in soil temperature (MacDonald *et al.*, 1995; Zogget *al.*, 1999). Thus these activities will influence the nitrification and denitrification processes, and thereby biotic formation of N₂O (Keeney *et al.*, 1979; Malhi *et al.*, 1990). Soil temperature is also responsible for the rate of gas diffusion from soils and influences the N₂O/N₂ ratio. This ratio increases with the decrease in soil temperature (Van Cleemput, 1998). For both temperate and tropical forest soils most researchers found a strong positive relationship between soil temperature and magnitude of N₂O emission (Gasche and Papen, 1999; Papen and Butterbach-Bahl, 1999; Sitaula and Bakken, 1993). However, Bowden *et al.* (1991) and Johansson *et al.* (1988) reported that N₂O emission was independent from changes in soil temperature indicating that other environmental factors, e.g. soil moisture, were stronger modulators for N₂O than soil temperature.

Soil temperature could be the main determinants for N₂O emissions provided that other factors are not limiting. This is particularly true for the temperate and humid tropics regions, where marked variations of temperature exist. However, a contrasting situation prevails in the humid tropics, where temperature ranges are identical and well distributed throughout the year. For example, Inubushi *et al.* (2003) investigated seasonal changes in N₂O emission over a whole year in tropical peatland in South Kalimantan. N₂O emission from abandoned agricultural land and secondary forest were low (ranging from -40 to 30 µgN m⁻² h⁻¹), and they found no clear seasonal changes in N₂O emission. They explained this result as inhibition of N₂O emission by flooding water.

In previously studied, Papen and Butterbach-Bahl (1999) and Gasche and Papen (1999) have demonstrated that the effects of soil temperature and soil moisture are not independent

of each other. They have shown that the best correlations between soil temperatures and N₂O emissions was found when water filled pore space (WFPS) was in an optimal range. McHale *et al.* (1998) had done a temperature simulation experiments which revealed that soil water content was a stronger modulator of N₂O flux than soil temperature. Similarly, most researchers had found that N₂O flux in temperate, boreal and tropical forest ecosystems is positively correlated to precipitation and WFPS (Goodroad and Keeney, 1984; Keller and Reiners, 1994; Riley and Vitousek, 1995; Sitaula and Bakken, 1993; Verchot *et al.*, 1999). However, there is a marked difference between forest types, whereby a beech stand explains 58% of the variation but a spruce stands only explains 4.7% of the variation (Papen and Butterbach-Bahl, 1999).

Soil pH has a significant influence on N transformation processes and proton (H⁺) budget of soils (Binkley and Richter, 1987; Van Miegroet and Cole, 1985). A negative correlation was found between soil pH value and magnitude of N₂O emission, i.e. decreasing soil pH seems to increase N₂O production (Fenn *et al.*, 1996; Kiese *et al.*, 2002). This may partially be caused by enhancing the chemical decomposition of the nitrite accumulated during NH₄⁺ oxidation (Venterea and Rolston, 2000). Acid peaty soils emit large amount of N₂O possibly due to pH induced changes in the N₂O:N₂ ratio (Flessa *et al.*, 1998) and in strongly acid soils, N₂O is the dominant end product during denitrification (Koskinen and Keeney, 1982; Parkin *et al.*, 1985). The toxic H⁺ and nutrient deficiencies at low pH might have reduce the vitality of the denitrifying microbes (Zech *et al.*, 1997).

There are conflicting results reported in the literature regarding the effect of soil pH on magnitude of both in situ N₂O from soils (Stevens *et al.*, 1998). These differences seem to be related to whether the pH value of the soil investigated reflects the natural development of the soil or the soil pH is changed by anthropogenic manipulations (e.g. liming and fertilization). The soil microbial communities may have adapted to environmental changes by selection of populations with a different pH or optimal temperature (De Boer *et al.*, 1995; Gødde and Conrad, 1999; Parkin *et al.*, 1985; Šimek and Hopkins, 1999). Thus prediction of the net effect may be problematic because production and consumption processes may be stimulated to varying extents.

NH₄⁺ and NO₃⁻ are the key substrates for nitrification and denitrification, the soil microbial processes responsible for production and emission of N₂O. Thus a strong positive correlation between soil NH₄⁺ and NO₃⁻ concentrations and magnitude of N₂O flux should be expected. However, even though most researchers describe such a relationship (Erickson

et al., 2001; Keller and Reiners, 1994; MacDonald *et al.*, 1997; Matson and Vitousek, 1990; Skiba *et al.*, 1998), there are other reports that could not detect any relationship between these parameters (Bowden *et al.*, 1991). Thus it is noted that in which a positive relationship between inorganic N and N₂O flux was detected, this relationship, though significant, was weak (Ambus and Christensen, 1995; Bowden *et al.*, 1991; Gasche and Papen, 1999; Matson and Vitousek, 1990; Papen and Butterbach-Bahl, 1999). These results indicate that the actual soil pools of NH₄⁺ and NO₃⁻ might be a minor significance for the prediction of N trace gas fluxes (Skiba *et al.*, 1998) and that actual microbial N turnover rates might be better predictors of in situ N₂O flux (Gasche and Papen, 1999; Papen and Butterbach-Bahl, 1999). A conceptual model has been proposed in which various environmental regulators affect ammonium (NH₄⁺) oxidation and NO₃⁻ reduction in a hierarchical way (Robertson *et al.*, 1989; Tiedje, 1988).

2.5.3.2 Biotic factors and human-induced influences

The main N₂O production and consumption processes must be located in the upper layers of soil, the humus layer where the supply of energy substrates is plentiful for microbial activities (Seiler and Conrad, 1981; Slemr *et al.*, 1984). In the temperate coniferous forest soils, the uppermost organic layer is the soil horizon contributing the most N₂O emission into the atmosphere (Borken and Brumme, 1997; Gasche and Papen, 1999; Goodroad and Keeney, 1985; Papen and Butterbach-Bahl, 1999; Skiba *et al.*, 1993). But in the deciduous beech forest, the mineral soil layer was the main contributor of about 78% of the N₂O emission (Papen and Butterbach-Bahl, 1999). Also for tropical rain forest in Australia, the mineral soil was identified to represent the most important soil layer for N₂O release into the atmosphere (Breuer *et al.*, 2000).

Vegetation type on identical soil type and exposed to identical climatic conditions is also a strong modulator for the magnitude of N₂O flux, because it influences the litter quality (i.e. C/N ratio of the litter) (Erickson *et al.*, 2001; Gasche and Papen, 1999; Menyailo and Huwe, 1999; Papen and Butterbach-Bahl, 1999). Forest site in the tropics with leguminous trees showed a 100-fold higher N₂O emission rates than sites lacking these N fixing tree species (Erickson *et al.*, 2001).

The enhanced N₂O emissions from natural and agricultural ecosystems are believed to be caused by increasing soil N availability driven by increased fertilizer application, agricultural N₂ fixation, and N deposition (Aber *et al.*, 2003; Galloway *et al.*, 2003; Hall and

Matson, 1999; Matson *et al.*, 1999; Skiba *et al.*, 1998). Application of different N sources under conventional and a sustainable land management system is the main source of N₂O. However, animal production systems, biological N fixation and erosion and leaching from soil systems are also the important contributors to N₂O emission (Galloway *et al.*, 2003; Skiba *et al.*, 1998). Drainage alone increases the N₂O emissions dramatically, but if the drained soil is used for agriculture, fertilization enhance the N₂O emission even more (Kasimir-Klemedtsson *et al.*, 1997). Fertilization and the form of N fertilizer affect the N₂O emissions during the month following fertilizer application (Flessa *et al.*, 1998; Skiba *et al.*, 1994), but weather after fertilization also affects the magnitude of the loss of fertilizer-N (Flessa *et al.*, 1998; Smith *et al.*, 1998). Although the effect of the chemical nature of the N fertilizers on denitrification and N₂O emissions is still subject of study (Eichner, 1990; Stehfest and Bouwman, 2006), it has been observed that NO₃⁻ based fertilizers may lead to high N losses from predominantly anaerobic soils (Scheer *et al.*, 2008), whereas the application of ammonium sulfate to aerobic soils can increase N₂O emissions up to 25.7 times (Trujillo-Tapia *et al.*, 2008).

2.5.4 Nitrification and denitrification for N₂O dynamics

The N₂O emission is a significant biogenic phenomenon in N transformation mechanism and occurs during both nitrification and denitrification process. They are the most important microbial processes in soils contributing to N₂O flux. Both denitrification and nitrification processes form and release N₂O, which are regulated by physical, chemical and microbiological conditions (Davidson, 1992; Firestone and Davidson, 1989; Payne, 1981, Poth and Focht, 1985; Sahrawat and Keeney, 1986; Yoshida and Alexander, 1970).

N₂O formation in soils is predominantly via denitrification in anaerobic microsites, brought about by the inhibition of aeration at high water contents (Davidson and Verchot, 2000; Linn and Doran, 1984; Smith *et al.*, 1998). The N₂O is an obligatory intermediate during denitrification. Denitrification is the sequential reduction of NO₃⁻ → NO₂⁻ → NO → N₂O → N₂, which is driven by the oxidation of organic or inorganic substrates (electron donors). Denitrifiers, especially organotrophs are aerobic microorganism, which can switch to anaerobic denitrification to consume C substrate in the absence of O₂ (Tiedje, 1988). Some denitrifiers can denitrify even in the presence of O₂ (Robertson and Kuenen, 1988). If soils containing NO₃⁻ become anaerobic, the availability of organic C that enhances the activity of denitrifiers is the limiting factor the reduction of NO₃⁻.

Nitrification may be a significant source of N₂O through autotrophic microbes in most soils and heterotrophs in aerobic to near aerobic soils (Anderson *et al.*, 1993; Bremner, 1997). Autotrophic nitrification is an aerobic process, which uses CO₂ as the principal C source for biomass formation. N₂O production is enhanced in soils having a high mineralization capacity to form NH₄⁺ or treated with nitrifiable forms of N.

In most forest soils, the importance of nitrification or denitrification as the main source of N₂O is not static. It can switch very rapidly, as the soil aeration state within the biologically active site changes due to rainfall or increased O₂ demand caused by the presence of easily mineralizable organic matter. Though denitrification has been demonstrated to be a very important source of N₂O from many forest soils (Davidson *et al.*, 1993; Firestone and Davidson, 1989; Keller and Reiners, 1994; Keller *et al.*, 1988), there is now increasing evidence that nitrification is the dominating process of N₂O emission from well aerated temperate and tropical soils during most parts of the year (Castro *et al.*, 1993; Erickson *et al.*, 2001; Matson and Vitousek, 1990; Papen and Butterbach-Bahl, 1999; Sitaula and Bakken, 1993). Although most of the N₂O in soil seems to be produced during NH₄⁺ oxidation and NO₃⁻ reduction, the exact production mechanism is usually not easily understood for a specific field situation.

Release of N₂O from the soil into the atmosphere is even more complicated, since consumption processes also take place and take control of the flux. It is obvious that production and consumption processes are regulated by environmental variables that affect the metabolism of the soil microorganisms involved. The production and consumption of N₂O by denitrification and nitrification in soils has been well described by a “hole-in-the-pipe” model as shown in **Fig. 2.7** (Davidson *et al.*, 2000). This model is used to describe two levels of regulation of N₂O emissions between the soil and atmosphere:

1. The amount of N flowing through the pipe, i.e. the rate of N cycling in general, or specifically to rates of NH₄⁺ oxidation by nitrifying bacteria and NO₃⁻ reduction by denitrifying bacteria; and
2. The size of the holes in the pipe through which gaseous N₂O, NO, and N₂ leak (or the amount of N that leaks out of the pipe) which is determined by several soil properties, but primarily regulated by soil water content. This effect of soil water content, and in some cases acidity or other soil factors, determine the relative rates of nitrification and denitrification and, hence, the relative proportions of gaseous end products of these processes.

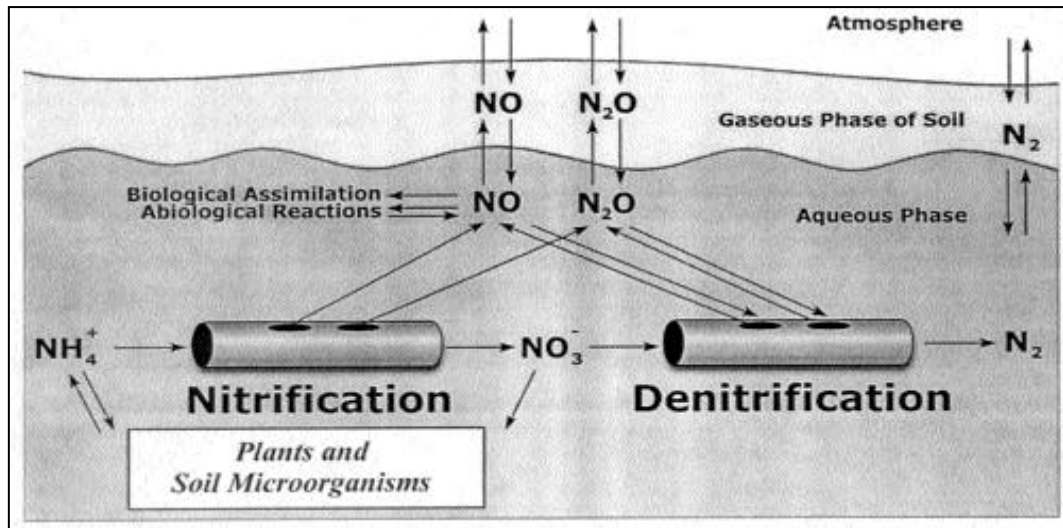


Fig. 2.7 Diagram of the “hole-in-the-pipe” conceptual model describing soil emissions of N₂O and NO (Davidson *et al.*, 2000).

The first level of regulation determines the total amount of N oxides produced (NO + N₂O) while the second level of regulation determines the relative importance of NO and N₂O as the gaseous end product of these processes.

2.6 Global warming potential (GWP)

GHGs are emitted in vastly different amounts and have widely different atmospheric lifetimes. The control options among them are also completely different. Therefore, GWP was developed as a single index for quantitatively comparing the climatic effects of equal emissions of the different GHGs. GWP can be used to compare the ability of each GHG to trap heat in the atmosphere relative to another gas. It also provides a means for comparing the relative effects of one source or sinks of GHGs against another and combines the warming effects of the individual GHGs to support the development of policies and measures design to mitigate climate change.

In general, only three GHGs are accounted for GWP: CO₂, CH₄ and N₂O. GWP values for emissions of CO₂, CH₄, and N₂O on a molar basis for a 100-year time horizon are 1:25:298. Although CH₄ and especially N₂O are far lower atmospheric concentrations than CO₂, their GWPs are so high that small changes have a disproportionate effect on radiative forcing. The GWP was used within the Kyoto Protocol to the United Nations Framework Convention on Climate Change as a metric for weighting the climatic impact of emissions of different GHGs (Oberthür and Ott, 1999). The GWPs vary depending on the time horizon

used. But in an effort to standardize the approach, the Kyoto Protocol uses a 100 year GWPs, which was recommended by IPCC for policy making and reporting purposes (IPCC, 1996).

The impact of climatically relevant emissions can be regarded, in a simplified manner, as the chain: emission changes → concentration changes → radiative forcing → climate impacts → societal and ecosystem impacts → economic “damage” (O’Neill, 2000; Smith and Wigley, 2000; Fuglestedt *et al.*, 2003). It has been recognized that, in general, the relevance of the impacts becomes greater as we move down this chain, and hence it would be desirable to have a metric that compares an impact that is more relevant than the time-integrated radiative forcing. However, it has also been recognized that uncertainty generally becomes greater as we move down this chain.

2.7 GHG concentrations dissolved in peat pore water

Numerous studies (e.g. Moore *et al.*, 1990a and b; Buttler *et al.*, 1991; Nilsson and Bohlin, 1993) have indicated high concentrations of dissolved CO₂ and CH₄ at depth in boreal and temperate peatlands. Previous studies have shown that under the field conditions CO₂ and CH₄ concentrations in the peat are mostly relatively uniform at levels of 1–10 mmol L⁻¹ for CO₂ and 0.1–1 mmol L⁻¹ for CH₄ (Nilsson and Bohlin, 1993). In addition, Blodau and Moore (2003) also reported CH₄ concentration in pore water at an oligotrophic peatland near Ottawa was generally ranged 10–190 μmol L⁻¹ just below the water table to 500–800 μmol L⁻¹ at depths of 40 to 70 cm. Dissolved CO₂ and CH₄ concentrations have also been proposed as a potential integrative measure of whole catchment soil respiration (Jones and Mulholland, 1998), particularly since both gases can be measured easily and directly (Kling *et al.*, 1991). These suggested the production, consumption, storage, and transport of gases in the peat was correlated with environmental variables.

In wetland soils, there is a close linkage between soil CO₂ fluxes and dissolved CO₂ concentrations. The dissolved CO₂ concentrations can act as an important “barometer” of the balance between soil respiration and photosynthesis. Clymo and Pearce (1995) have explained that the diffusive movement of soil gases through waterlogged peat is slow ~315 cm² yr⁻¹. In the studies of physiology and cell biochemistry, diffusion may be effective in transport only for distances up to a few micrometers. In field conditions, however, the distances are five orders of magnitude greater (Clymo and Bryant, 2008). On the other hand, soils can be more freely draining and aerated, the proportion of soil respired CO₂ lost from the soil surface is greater and hence the role of water as a conduit for gas escape and as a barometer of soil CO₂ is decreased.

The spatial and temporal variability of dissolved CH₄ is large. Microbial processes (methanogenesis and CH₄ oxidation) play a major role in the release of CH₄ to the atmosphere. The high concentrations of CH₄ in shallow soil are particularly common where standing water is present, because it impedes entry of O₂ to support methanotrophy (Bartlett *et al.*, 1988; Koschorreck, 2000). Svensson and Rosswall (1984) reported the variations in dissolved CH₄ profiles in acidic peat soils; CH₄ concentrations were higher at minerotrophic sites in which the water table was at the vegetation surface compared to ombrotrophic sites in which the water table was 5–6 cm below the surface. Dise (1993) reported seasonal variations in the vertical distribution of CH₄ (10 cm depth intervals) in bogs, but because of the indirect method employed, high spatial resolution was not achieved. Thus, Sebacher *et al.* (1985) observed a peak in the vertical distribution of dissolved CH₄ in sediment pore water near the rhizomes of aquatic plants.

The concentrations of N₂O in soil solutions are also of very high temporal and spatial variability (Bowden and Bormann, 1986; Dowdell *et al.*, 1979; Minami and Fukushima, 1984; Minami and Oshawa, 1990; Ronen *et al.*, 1988; Schnabel and Stout, 1994; Terry *et al.*, 1981). Transitorily significant quantities of N₂O can be dissolved in the soil solution. The maximal concentrations of dissolved N₂O reported are up to three orders of magnitude above ambient. On the other hand, temporary N₂O under saturation is also possible (Bowden and Bormann, 1986; Davidson and Firestone, 1988; Terry *et al.*, 1981). Similar to the horizontal, the vertical distribution of N₂O in soils is very different. In some studies the concentrations of N₂O in the soil atmosphere were equal at all soil depths (Albrecht *et al.*, 1970; Parkin and Meisinger, 1989). In other cases the spatial variability was in general so high that a distinction between layers was impossible (Cates and Keeney, 1987). The high temporal variability of N₂O concentrations in the soil profiles is in part the result of seasonal and chemical factors regulating N₂O dynamics. Some investigations revealed seasonal dynamics of N₂O concentrations in the soil solution (Bowden and Bormann, 1986; Minami and Fukushima, 1984). Other reports, however could not confirm such systematic relations (e.g. Minami and Oshawa, 1990).

Few investigations take the transport of dissolved N₂O with the soil solution and further out-gassing into the atmosphere. Leaching of dissolved N₂O requires high N₂O production rates and great amounts of infiltrating water causing fast preferential flow (Dowdell *et al.*, 1979; Schnabel and Stout, 1994). In the relation of N₂O production, the denitrification being a major N₂O production process, clear relationships between increasing concentrations of N₂O and decreasing availability of O₂ in the soil air have been often shown

(Burford *et al.*, 1981; Dowdell and Smith, 1974; Eggington and Smith, 1986; Hansen *et al.*, 1993; Parkin and Meisinger, 1989; Rolston *et al.*, 1976). In wet soils the N₂O flux increased four times more slowly with rising N₂O concentrations in soil air than in dry soils due to a much smaller diffusion coefficient. Some studies found no or only weak positive correlations of the N₂O concentrations in the soil air and the N₂O emissions (Ball *et al.*, 1997; Clayton *et al.*, 1994; Goodroad and Keeney, 1985; Lessard *et al.*, 1996; Velthof *et al.*, 1996). There may be another explanation for these discrepancies. The main N₂O production may be take place either in the upper few centimeters of the soil with rapid diffusion out of it, or too deep in the soil profile causing a time lag between production and emission of N₂O (Goodroad and Keeney, 1985). Other factors may be large spatial variability of N₂O concentrations and N₂O fluxes or the dissolution of N₂O in the soil solution (Velthof *et al.*, 1996). Furthermore, high water contents may restrict N₂O diffusion and enforce the microbial reduction of N₂O to N₂ (Lessard *et al.*, 1996).

There have been several studies on the dissolved GHG concentrations in tropical peatlands. Ueda *et al.* (2000) reported the concentrations of dissolved gases in a coastal swamp system in Thailand varied widely in swamp water, river water, and ground waters: 1–3290 $\mu\text{molCO}_2 \text{ L}^{-1}$; 0.01–417 $\mu\text{molCH}_4 \text{ L}^{-1}$; 0–3.3 $\mu\text{molN}_2\text{O L}^{-1}$. Pangala *et al.* (2013) reported the amount of CH₄ in deeper peat in Borneo peatland was greater (113–1539 $\mu\text{mol CH}_4 \text{ L}^{-1}$) at depths of 50–150 cm. However, ~83% of root biomass occurs within 0–30 cm depths in the tropical peat forest and root abundance decrease exponentially with depth (Jauhiainen *et al.*, 2005; Sulistiyanto *et al.*, 2004; Verwer and van der Meer, 2010). In contrast, shallow soil (0–30 cm) in more nutrient-rich tropical wetlands (Amazonian wetlands) has been reported to contain dissolved CH₄ concentrations of 175–1380 $\mu\text{molCH}_4 \text{ L}^{-1}$ (Bartlett *et al.*, 1988; Koschorreck, 2000).

2.8 DO concentrations in peat pore water

DO concentration in tropical peatland soils plays a major role. It regulates or controls the microorganism activities, which affect the decomposition processes of organic matter, and hence indirectly affect the GHG emissions from soils to the atmosphere. DO availability also has controlling influences on the key aspects of wetland biogeochemical cycles by determining the relative rates of aerobic to anaerobic metabolism (e.g. Patrick and Reddy, 1976).

The wide spread assumption that water table level drawdown stimulates peatland microbial activity has been investigated. The origins of the assumption are difficult to

determine, although it seems likely to have been based on two premises: 1) O₂ diffusion through waterlogged peat is *ca.* 10⁻⁴ of that in air (Clymo, 1983), and 2) aerobic metabolism is considerably more efficient than its anaerobic counterpart (Stainer *et al.*, 1978). Moreover both in laboratory and field studies indicated that drier conditions stimulate mineralization (Freeman *et al.*, 1993c; Heathwaite, 1990). The concentration of DO is low due to saturated conditions of peatlands. Under low DO concentrations, the rates of decomposition processes become slow, meanwhile this condition is favorable for the production of CH₄ and N₂O (by denitrification) in soils. In contrast, activity of microorganism will increase under higher DO levels accompanied by lower water table levels and the entrance of O₂.

Though O₂ has long been recognized as an important driver for determining microbial metabolism, relatively little is known about field-level O₂ dynamics in soils (Burgin *et al.*, 2010; Liptzin *et al.*, 2011; Silver *et al.*, 1999). Consequently, our understanding of how soil O₂ varies is limited. This restricts our understanding of how soil O₂ variation affects ecosystem processes including GHG production (Whalen, 2005), nutrient dynamics (Crawford, 1992), and redox (Pett-Ridge and Firestone, 2005). In the aquatic marine ecosystem, DO is usually used to express the toxic conditions for marine ecosystems. DO concentration also affects the key aspects of major biogeochemical cycles, thereby altering the bioavailability of both nutrients and toxic compounds (Reddy and D'Angelo, 1994). The principles of DO processes in marine ecosystems can be applied to the wetland ecosystems.

The subsurface O₂ concentrations in wetlands have also rarely been reported at high spatio-temporal scales. For example, Ueda *et al.* (2000) reported that DO concentrations in swamp water, river water, and ground water at coastal swamp system in Thailand was 0–0.248 mmol L⁻¹. DO has been proposed as a key parameter for consideration in wetland management and restoration programs (Yozzo and Titre, 1996), but is often excluded from wetland monitoring programs (e.g. Brooks and Hughes, 1988; Burgin and Groffman, 2012).

While few studies have measured soil O₂, it was dynamic under varying conditions in water table (Burgin *et al.*, 2010), seasonal water table dynamics (Faulkner and Patrick, 1992; Megonigal *et al.*, 1993) or precipitation (Liptzin *et al.*, 2011; Silver *et al.*, 1999). Previous studies also reported that wetlands with highly colored water are characterized by low water-column DO and show little daily variation (Belanger *et al.*, 1985; Hampson, 1989).

Some studies found the relationship between distribution of plant roots and oxygen concentration in subsurface soils (Burgin and Groffman, 2012; Ding *et al.*, 2004; Elberling *et al.*, 2011; Fritz *et al.*, 2011; Joabsson *et al.*, 1999; Pangala *et al.*, 2013; Pezeshki, 2001; Whalen, 2005; van der Gon and Neue, 1996). In the peatland forest, trees have the capacity to

cope with soil anoxia through development of morphological adaptations such as hypertrophied lenticels, adventitious roots and enlarged aerenchyma. These structures promote gas exchange between the atmosphere and the rhizosphere (Kozłowski, 1997; Megonigal and Day, 1992). In addition gas transport through herbaceous plants adapted to wet soil is well documented (Brix *et al.*, 1992; Whiting and Chanton, 1996), in particular, entry of O₂ to the roots zone.

Table 2.1 Comparison of the fluxes and dissolved concentrations of CO₂, CH₄, and N₂O in wetland ecosystems in the world.

No.	Location	Land use	Peatland/soil type	Sampling period	Flux			Dissolved concentration			References
					CO ₂ (mgC m ⁻² h ⁻¹)	CH ₄ (mgC m ⁻² h ⁻¹)	N ₂ O (µgN m ⁻² h ⁻¹)	CO ₂ (mmol L ⁻¹)	CH ₄ (mmol L ⁻¹)	N ₂ O (µmol L ⁻¹)	
1	Alpine wetland ecosystem in Luanhaizi, Qinghai-Tibetan Plateau, China 37° 29' N, 101° 12' E	Alpine meadow ecosystem Emergent-plant		July 2002 to September 2002	Mean 2.46						Hirota <i>et al.</i> , 2004.
2	Peatland near St. Charles-de-Bellechasse, Quebec, Canada 46° 40' N, 71° 10' W	Hummock, lawn, and hollow Forest	Fen fragment	May 2002 and January 2003							Strack <i>et al.</i> , 2004.
		Drained site									Mean 1.4 to 5.8 (2002)
											Mean 0.3 to 0.7 (2002)
3	Peatland in the Mer Bleue, near Ottawa in eastern Ontario, Canada	<i>Vegetations of mosses</i>	Oligotrophic peatland	June 2003 to December 2004				0.1-2.5	0-0.6		Blodau <i>et al.</i> , 2007.
4	The Mer Bleue site near Ottawa, eastern Ontario, Canada	Mosses and shrubs	Oligotrophic peatland						0.01-0.190		Blodau and Moore, 2003.
5	Peatland in Stor-Vidmyran in northern Sweden	<i>Sphagnum peat</i>	Mire peatland	in 1989 and 1990				0.97-6.5	0.05-1.2		Nilsson and Bohlin, 1993.
6	Bog lake peatland in the Chippewa national forest in northcentral Minnesota 47° 32' N, 93° 28' W	Forest, <i>Sphagnum</i> vegetations	Minerotrophic to oligotrophic	1991 and 1992 (growing season)	2.50-5.50						Shurpali and Verma, 1998.
7	The Sanjiang Mire Wetland Experimental Station, Chinese Academy of Sciences, in Tongjiang city in the eastern part of Heilongjiang province, Northeast China 47° 35' N, 133° 31' E	Vegetation varies: <i>Carex lasiocarpa</i> <i>Carex meyeriana</i> <i>Deyeuxia angustifolia</i>		August 2001		21.6-66.5			24-229		Ding <i>et al.</i> , 2004.
									17.2- 45.0		8-202
									19.1-33.0		54-232

Table 2.1 continued

8	The Wildmooswald in the cool-humid Central Black Forest in Germany 47° 57' N, 8° 07' E	Mature Norway spruce (<i>Picea abies</i> L.)	Haplic Gleysol Histic Gleyson Sapric Histosol	47-129 60-158 65-172	-0.024-0.072 -0.012- 0.094 0.010-0.141	81-449 83-963 20-653	Jungkunst <i>et al.</i> , 2008.
9	The north of Great Hing'an Mountains, northeast China 52.94° N, 122.86° E	Vascular plants	Minerotrophic peatland	June 2010 to September 2011	0.21-1.02	Mean 0.01437	Miao <i>et al.</i> , 2012.
10	The Asa experimental forest, in southern Sweden 57° 08' N, 14° 45' E	Drained site and undrained site	Sedge mire	August 1999 to September 2002	Median 70 to 1200	Median 1 to 4	Von Arnold <i>et al.</i> , 2005.
11	The east and southwest of the Netherlands	Lakes Ditches	Eutrophic to oligotrophic	June to July (Summer) 2009	1.4-18.1		Schrier-Uijl <i>et al.</i> , 2011.
12	Drainage catchment area, near the headwater of a tributary of Ythan River catchment, Aberdeenshire, UK	Pasture farm (mixed arable/grassland) <i>Field drain</i> <i>Open farm yard ditch</i>	Podzols, brown forest soil, and noncalcareous gleys soil	August 29, 2002	69.6-199.0	1.2-39.3	Reay <i>et al.</i> , 2004. Mean 28.2 Mean 8.5
13	The Pärnu Lowland in western part of Estonia 58° 34' 20" N, 24° 23' 15" E	<i>Open ditch</i> Combination of peat extraction: <i>Bare peat</i> <i>Nonfertilized Phalaris site</i> <i>Fertilized Phalaris site</i> <i>Natural raised bog site</i>	Mineralized fibric histosol	May 2010 to May 2011	-0.021-0.9	-1.9-261	Mander <i>et al.</i> , 2012.
14	The Lompolojänkkä northern boreal fen at north-western Finland in the aapa mire region 67° 59.932' N, 24° 12.551' E	Mosses, sedges, and shrubs	Nutrient-rich sedge fen	September 2004 to October 2008		-45-37	Lohila <i>et al.</i> , 2010.

Table 2.1 continued

15	Peat swamp basin in Mukah Division of Sarawak Malaysia 2° 49' N, 111° 51' E to 2° 49' N, 111° 56' E	Forest Sago Oil palm	Typic tropofibrists	November 2002 to May 2003	100-533 63-245 46-335	Melling <i>et al.</i> , 2005a.
16	Peat swamp basin in Mukah Division of Sarawak Malaysia 2° 49' N, 111° 51' E to 2° 49' N, 111° 56' E	Forest Sago Oil palm	Typic tropofibrists	November 2002 to May 2003	-0.000453-0.0084 -0.00744-0.10206 -0.03278-0.00417	Melling <i>et al.</i> , 2005b.
17	Peat swamp basin in Mukah Division of Sarawak Malaysia 2° 49' N, 111° 51' E to 2° 49' N, 111° 56' E	Forest Sago Oil palm	Typic tropofibrists	November 2002 to May 2003	-3.4-19.7 1.0-176.3 0.9-58.4	Melling <i>et al.</i> , 2007.
18	Tropical peatland, upper Sebangau River Southeast of Palangka Raya, Central Kalimantan 02° 20' S, 113° 55' E	Forest <i>Hollows</i> <i>Hummocks</i>	Typic tropofibrists	in 1999, 2000, and 2001 during wet and dry season	Mean 139 Mean 484	Jauhainen <i>et al.</i> , 2005.
19	Tropical peatland, upper Sebangau River Southeast of Palangka Raya, Central Kalimantan	Forest <i>Hollows</i> <i>Hummocks</i> Stem of plants	Typic tropofibrists	March 2011	Mean 0.0329 Mean 0.0007 0.017-0.185	Pangala <i>et al.</i> , 2013. 0.113-1.539 (50 - 150 cm depths)

Table 2.1 continued

	Forest sites:	TypicTropofibrists	December 2011 to May 2012									
20	Tropical peatland near Palangka Raya city, Central Kalimantan, Indonesia 02° 19' 25.20" S, 113° 54' 16.86" E			Mean 186	Mean 0.424	Mean -1.24	Mean 0.698 (all depths)	Mean 0.00147 (all depths)	Mean 0.0040 (all depths)	Present study		
		<i>FW1</i>										
		<i>FW2</i>	Mean 200	Mean 1.93	Mean -3.1	Mean 0.659 (all depths)	Mean 0.00082 (all depths)	Mean 0.0007 (all depths)	Present study			
		<i>FD</i>	Mean 340	Mean 0.008	Mean 3.4	Mean 2.03 (all depths)	Mean 0.00086 (all depths)	Mean 0.10 (all depths)	Present study			
		Burnt sites:										
		<i>BW1</i>	Mean 353	Mean 8.32	Mean -1.5	Mean 1.24 (all depths)	Mean 0.12 (all depths)	Mean 0.005 (all depths)	Present study			
		<i>BW2</i>	Mean 120	Mean 4.46	Mean -11.1	Mean 1.10 (all depths)	Mean 0.129 (all depths)	Mean 0.02 (all depths)	Present study			
		<i>BD</i>	Mean 108	Mean 0.220	Mean 8.1	Mean 0.836 (all depths)	Mean 0.0452 (all depths)	Mean 0.006 (all depths)	Present study			
21	Kalampangan village near Palangka Raya Central Kalimantan, Indonesia2° S, 114° E			Mean 5.0 (dry season) and 49 (wet season)	Mean 5.0 (dry season) and 49 (wet season)					Takakai <i>et al.</i> , 2006.		
		Natural forest										
		Regenerated forest										
		Burnt forest										
		Grassland										
		Agriculture:										
		<i>Plot A</i>	Mean 163 (dry season) and 4298 (wet season)	Mean 163 (dry season) and 4298 (wet season)								

Table 2.1 continued

						Mean 154 (dry season) and 637 (wet season)		
						Mean 94 (dry season) and 2441 (wet season)		
	<i>Plot B</i>						0.175-1.380	Koschoreck, 2000.
	<i>Plot C</i>							
22	Peatland at Marchantaria Island, an island in Amazon River, near Manaus, Brazil 03° 13.947' S, 59° 56.765' W	Shallow part of lake Camaleão (Floodplain ecosystem)	Transition of the sandy plains and the swamps	Wet season (Nov-Dec, 1991-1993) and dry season (Jul. 1992 and Aug. 1993)	October to November 1996 and 1997			
23	The Daeng and Bacho swamp peatland near Narathiwat province in Thailand	Mangrove forest				0-3.290	0.00001-0.417	Ueda <i>et al.</i> , 2000.
23	Peatland in southernmost Patagonia, Tierra del Fuego 54° 45' S, 68° 20' W	Vascular plants	Bog peatland		Spring 2006 to summer 2007		Mean < 0.001 (rooting zone)	Fritz <i>et al.</i> , 2011.
24	Wolong Marsh and Tuanjie Marsh in the north of Milor Peninsula of the Larseman Hills, eastern Antarctica 69° 23' to 69° 56' S, 76° 20' to 76° 45' E	Vegetations of algae community <i>Lynobya</i> , <i>Nostoc</i> , <i>Cosmarium</i> , etc	Marsh (Wetland ecosystem)		December 2005 to February 2006	-20.6-85.6	Mean 0.166 (> 170 cm depth)	Zhu <i>et al.</i> , 2008.

DESCRIPTIONS OF STUDY SITE

3.1 Overview of Central Kalimantan

The Province of Central Kalimantan – Indonesia situated near the equator (tropic island) between latitudes 0° 45' N to 3° 30' S and longitudes 111° to 116° E. It occupies an area of 153,564 km² with a population of just 2.2 million (Badan Pusat Statistik – Kalimantan Tengah, 2012), having a very low population density. The capital of the Province is Palangka Raya. There are 13 local government districts and 1500 urban and rural villages. Overall 67% of the population live in rural areas, where villages tend to be distributed along the banks of the 11 large rivers which flow from the central mountains to the Java Sea. The largest ethnic group is the indigenous Dayaks, but there are significant numbers of official and economic migrants from Java, Bali, and Sulawesi in both rural and urban settlements (<http://www.kalteng.go.id>). According to the Wetlands International – Indonesia (2004), the extents of peat soil in Central Kalimantan – Indonesia is 3,010,640 ha (52.18%), a total distribution of peat soils in Kalimantan is 5,769,246 ha.

3.2 Study site

The field observation was conducted in a tropical peat swamp area located near Palangka Raya City (02° 19' 25.20" S, 113° 54' 16.86" E). Much area of the peat swamp forest in this region was deforested and drained during a national project in late 1990s, the Mega Rice Project (MRP). Although the project was terminated in 1999, it left vast tracts of devastated peatland (Muhammad and Rieley, 2002; Notohadiprawiro, 1998). A large portion of the devastated peatland was subsequently abandoned because it was unsuitable for the intended agricultural production. In these areas, several large fires have occurred during the dry season, severely damaging both the vegetation and soil. These fires were mainly ignited as a result of agricultural activities, including land clearance and ash production by local farmers. Some of these fires spread uncontrollably to become wildfires (Usup *et al.*, 2004).

Six locations were established as study sites (**Fig. 3.1**). Three sites were placed in peat swamp forest that was neither affected by deforestation or fire (FW1, FW2, and FD; forest sites). The other three sites were located in a degraded peatland and recently affected by fire (BW1, BW2, and BD; burnt sites). The burnt sites were deforested in 1995 and burned down in 1997, 2002, and 2009 (El Niño years). The flooded forest sites (FW1 and FW2) and the flooded burnt sites (BW1 and BW2) were close to the flood plains of Sebangau River and

Kahayan River, respectively. FD and BD sites were located near the top of a peat dome which is formed between these two rivers. These two sites were subject to be influenced by a large drainage canal, which was extended to the Java Sea (>100 km).

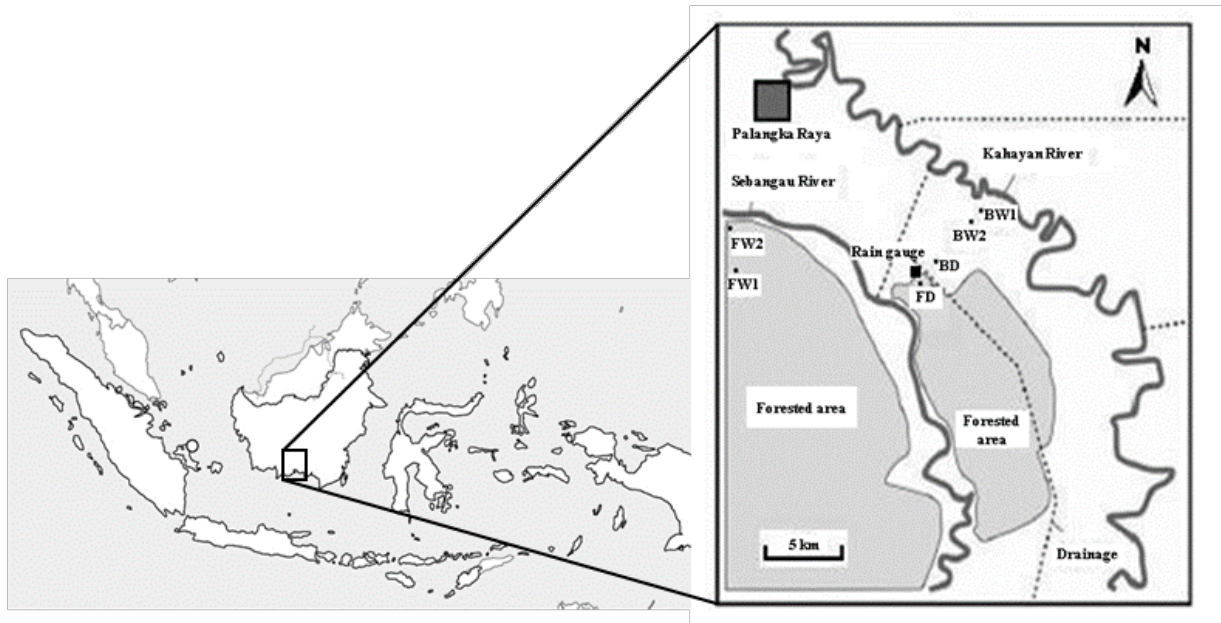


Fig. 3.1 Locations of field observation sites.

Annual precipitations and annual mean air temperature for the nine years of 2002–2010 were 2540 ± 596 mm and $26.2 \pm 0.3^\circ\text{C}$, respectively (Sundari *et al.*, 2012). Variation in annual precipitation was large; the maximum (3750 mm) and minimum (1852 mm) were recorded in a La Niña year (2010) and an El Niño year (2002), respectively. Air temperatures showed little seasonal and inter annual variations. The dry season normally begins in June and ends in October. The field observation was conducted during the wet season from December 2011 to May 2012, to simulate high water level conditions.

In the forest sites, the ground surface was covered by a dense canopy of peat swamp forest, although the composition of tree species was previously affected by selective logging until the late 1990s (Hirano *et al.*, 2012). Dominant species present included *Tristania whittiana*, *Combretocarpus rotundatus*, *Palaquium leiocarpum*, and *Stemonurus scorpioides*. In the burnt sites, the ground surface was covered by ferns and grasses, such as Kalakai (*Stenochlaena palustris*) and Pakis/Paku-pakuan (*Nephrolepis* sp.), with patchy depressions. Trees, such as Tumih (*Combretocarpus rotundatus*), were found sparsely. Detailed information about the vegetation around the study sites is provided

elsewhere (Jauhiainen *et al.*, 2005; Page *et al.*, 1999; Sundari *et al.*, 2012; Takakai *et al.*, 2006; Tuah *et al.*, 2001 and 2003).

Based on USDA Soil Taxonomy, the study sites were all classified as Typic Tropofibrists (Takakai *et al.*, 2006). The peat depths were 2–3 m at FW1 and FW2, and about 4 m at other sites. Volumetric carbon density of peat was $71.5 \pm 17.3 \text{ kg m}^{-3}$ (Shimada *et al.*, 2001).

ENVIRONMENTAL FACTORS ON TROPICAL PEAT DECOMPOSITION AND GHGs DYNAMICS

4.1 Introduction

The interaction of vegetation, hydrology, and soil results in the environmental characteristic unique to an ecosystem. In the greenhouse gas study, it is very important to have a good understanding of the environmental characteristics of an ecosystem because it highly influences the dynamics of GHGs in tropical peatland. Thus, this chapter outlines the findings of the environmental investigations conducted as part of the environmental study and provides details of environmental characteristics for each study sites.

4.2 Materials and methods

4.2.1 Precipitation and water table level

Precipitation was measured in the vicinity of the study sites (**Figs. 3.1 and 4.1**) with a tipping bucket rain gauge linked to a HOBO Pendant[®] event data logger. Water table level was measured and recorded with a pressure logger (Hobo U20, Onset, Bourne, USA) settled near the bottom of a perforated PVC pipe (ca. 1.5 m from the ground surface). The influence of the change in atmospheric pressure was canceled based on the pressure data observed at the ground surface level inside the well of FD, where the water table level was lower than -20 cm during the observation period (see **Results**). The water table level in the well was converted to that at the position of the flux chambers by canceling the small differences in the altitude of the ground surface.



Fig. 4.1 View of the tipping bucket rain gauge settled near the base camp of Kalampangan canal.

4.2.2 Air and soil temperatures

Air temperature was measured at a height of 100 cm with a digital thermometer (TESTO 625, Lenzkirch, Germany) during the flux chamber measurement (see **Chapter 5** in detail). Prior to the field observation in this study, hourly monitoring of soil temperature was conducted in February, June, and December 2011 at the flooded sites, and in July 2011 at all the study sites. Soil temperature was measured at depths of 5, 10, 30, and 50 cm with a T-type thermocouple and recorded to a 4-ch data logger (47SD Thermo Recorder, Sato Shoji, Japan). One monitoring campaign was usually continued 4 to 10 consecutive days.

4.2.3 DO Concentration

DO concentrations (mg L^{-1}) were measured in situ by extracting pore water directly into a closed cell equipped with a DO sensor to avoid contact between the water sample and the atmosphere (**Fig. 4.2**). A luminescent DO probe (LDO10101; Hach, Loveland, USA) was used because, in principle, it neither consumed DO nor required water flow toward the sensor during measurement. The closed cell consisted of the tip of the probe, a 60-mL plastic syringe, and a connecting PVC adapter (**Fig. 4.3**). Pore water was extracted from a water sampler consisting of a stainless steel pipe (0.7 cm in diameter), a plastic tube and a three-way

stopcock and separately installed at depths of 10, 20, 40, and 80 cm (**Fig. 4.4**). During the observation intervals, the sampler was filled with water and the stopcock was kept closed to prevent the entrance of the aboveground atmosphere. The water was first stored in the preparatory syringe and then pushed into the closed cell (**Fig. 4.3**). As the water entered, the cell plunger was synchronously pushed out. After measurement, the water was drained from another cell port. DO was measured in December 2011 and March, April, and May 2012.

Dead space in the closed cell was initially filled with ambient air (first measurement) or pore water at different depths (second or later measurement). To approximate the true value of the DO concentration in the cell, 50 mL of pore water was repeatedly extracted until the measured value was stabilized. The stroke of the extraction was limited to 10 times to reduce the influence of the inflow of pore water from other depths. After the first DO measurement, the remaining water in the closed cell was not drained to avoid the invasion of ambient air and the influence of atmospheric O₂, which was much higher than those dissolved in pore water.

Finally, DO was determined based on the sequential data of measured DO by solving a numerical convergence model with Excel[®] Solver, as followed the equations:

$$C_n = C^*a + C_{n-1}(1 - a) \quad (4.1)$$

$$C^* = [C_n - C_{n-1}(1 - a)]/a \quad (4.2)$$

where, C_0 is an initial concentration of DO ($C_0 \geq 0$), C_n is measured value of DO in n extraction of peat water ($C_n \geq 0$), C^* is convergent value of DO, namely the result of the calculation, and a is a virtual exchange ratio of water volume collected in a syringe chamber ($0 < a < 1$).



Fig. 4.2 View of DO measurement at the study site.

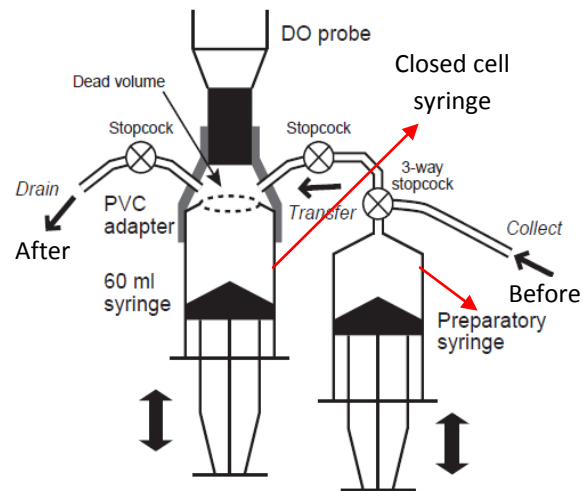


Fig. 4.3 Diagram of an in-situ DO measurement.

4.2.4 pH and EC

pH and electric conductivity (EC) were measured at depths of 10, 20, 40, and 80 cm (**Fig. 4.4**). Water samples were extracted from the same samplers for the DO measurement. The values of pH and EC were determined in situ with portable sensors for pH (B-212, Horiba, Japan) and EC (B-173, Horiba, Japan), respectively.

4.2.5 Dissolved nitrogenous ions

For the measurement of dissolved nitrogenous ions (NO_3^- and NH_4^+), 50 mL of water were collected to a plastic bottle at depths of 10, 20, 40, and 80 cm using the same samplers of DO (**Fig. 4.4**). These water samples were stored in a refrigerator until analysis. Concentration of NO_3^- was measured by ion chromatography (Dionex QIC Analyzer, Thermo Fisher Scientific, Waltham, USA). Concentration of NH_4^+ was determined by colorimetry based on the indophenol-blue method with an ultraviolet-visible spectrophotometer (UV mini 1240, Shimadzu, Kyoto, Japan).



Fig. 4.4 View of a set of water samplers for pH, EC, NO_3^- , NH_4^+ , and DO measurements at the study site.

4.2.6 Statistical analysis

A one-way analysis of variance (ANOVA) was conducted to test for differences in the environmental parameters among the six study sites. The effects of land use, drainage, and depth on DO concentration was tested using three-way ANOVA. Multiple comparison tests among the sites and depths were conducted using the Bonferroni method. A probability level less than 0.05 ($P < 0.05$) was considered statistically significant. All analyses were performed in IBM SPSS Statistics Ver. 21.

4.3 Results and discussion

4.3.1 Precipitation and water table level

Precipitation and water table level showed temporal changes during the observation period (Fig. 4.5). Water table level at all sites occasionally rose in response to high rainfall intensities, but it was generally stable from Dec. 2011 to Apr. 2012. In late May, the water table level was lowest due to the low monthly precipitation in May (94 mm, while 205 to 371 mm in the other months). Sites FW1, BW1, and BW2 generally remained flooded during the study period. At BW1, a temporary rise in water table level was recorded in late March despite the fact that no heavy rainfall was recorded and that no similar increases at other sites were observed. The rise in water table at BW1 may have been caused by an increase in the Kahayan River's water level. Sites FW2 and BD remained flooded for 45% and 60% of the observation period, respectively. With a water table level at below -20 cm during the study period, site FD was never flooded.

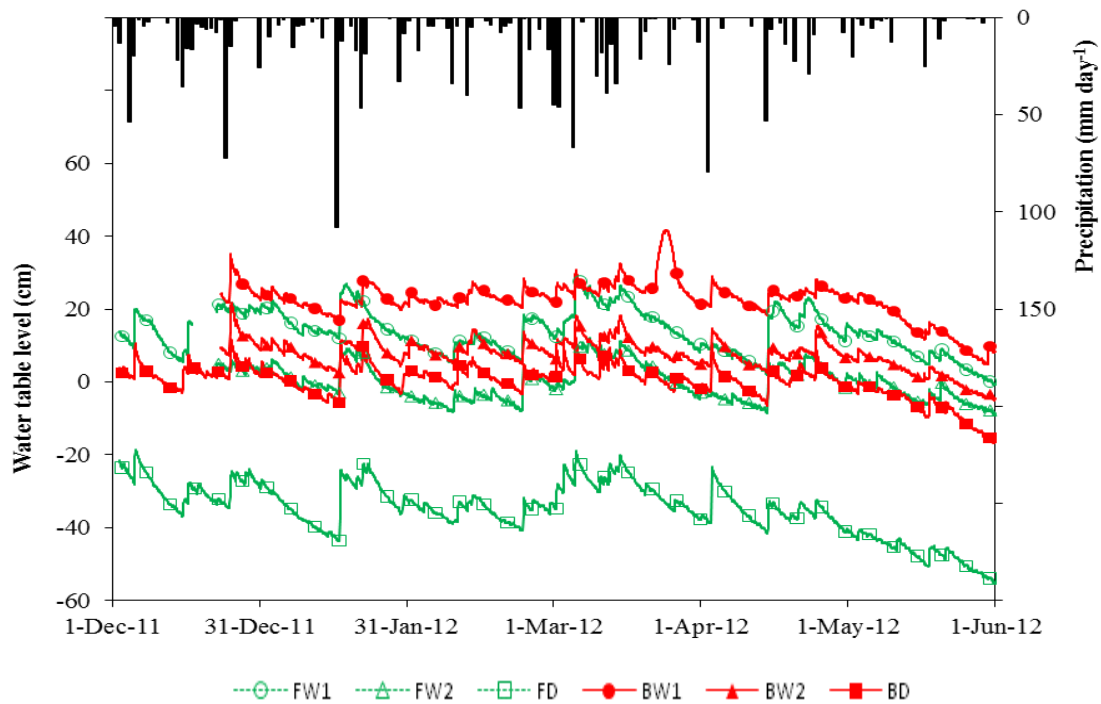


Fig. 4.5 Seasonal changes in precipitation (bar) and water table level at each site (line with symbol).

4.3.2 Air and soil temperatures

Air temperature observed during the flux chamber measurement was summarized in **Table 4.1** and **Fig. 4.6**. Air temperatures at the burnt sites (32–34°C) were higher than those at the forest sites (27–28°C), while there was no significant difference within each land use. This is because the forest sites were more shaded with dense vegetation. Diurnal variation in soil temperature reached up to 6°C at a depth of 5 cm, while less than 1°C at 50 cm. In spite of such large differences in diurnal variation, difference in daily mean soil temperature was small among the periods (February to December 2011; up to 3.3°C at 5 cm in BW1, <2°C at 10 to 50 cm in all sites) and depths.

Table 4.1 Air temperature at all study sites averaged for the observation period (Dec. 2011 to May 2012).

Site	Description	Air temperature (°C)
FW1	Flooded forest site #1	28.2 ± 1.0 ab (3)
FW2	Flooded forest site #2	27.4 ± 1.2 a (5)
FD	Drained forest site	27.4 ± 1.2 a (5)
BW1	Flooded burnt site #1	33.9 ± 4.0 c (2)
BW2	Flooded burnt site #2	31.7 ± 2.2 bc (4)
BD	Drained burnt site	32.8 ± 1.4 c (5)

1. Values are means and standard deviations of environmental factors.
2. Numbers in parentheses represent sample size at each site.
3. Values within the same column with different lowercase letters differ significantly among sites (Bonferroni, $P < 0.05$). The values increase in alphabetical order.

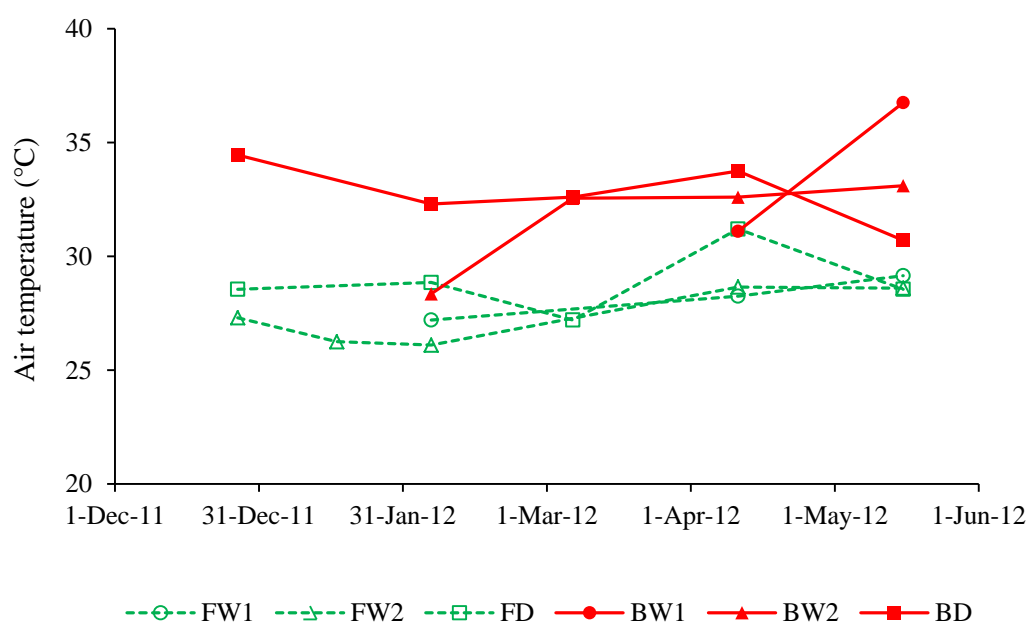


Fig. 4.6 Seasonal variations in air temperature at each study sites.

Vertical distributions of the average soil temperature are shown in **Fig. 4.7**. The average soil temperature was 24.5–25.5°C at the forest sites and 27.6–28.1°C at the burnt sites, reflecting the difference in air temperature.

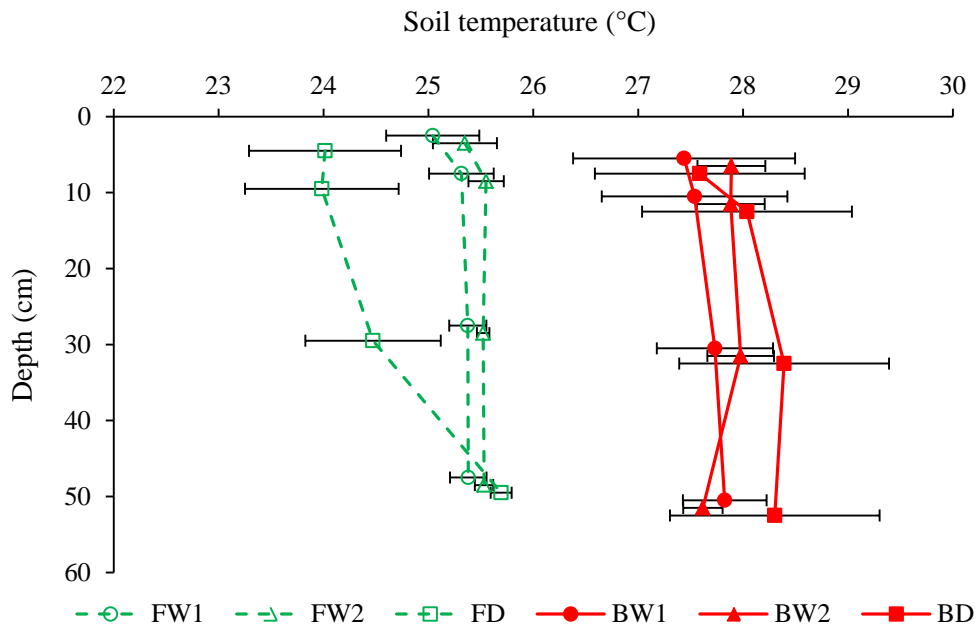


Fig. 4.7 Vertical distributions in soil temperature at all soil depths. Error bars show standard deviation. For better identifiability, the depths in the profiles were slightly shifted from their actual depths (5, 10, 30, and 50 cm).

4.3.3 DO concentration

DO concentrations ranged from 0.5 to 25 $\mu\text{mol L}^{-1}$, which is generally $<10 \mu\text{mol L}^{-1}$, much lower than DO in a solution which is in equilibrium with the atmospheric O_2 level (220–280 $\mu\text{mol L}^{-1}$), and showed little change with depth (**Table 4.2 and Fig. 4.8**). DO concentrations were occasionally changed at all sites (**Fig. 4.9**). According to the results of three-way ANOVA, only the main effect of land use was significant ($P < 0.05$; forest $>$ burnt, 6.9 ± 5.6 and $4.0 \pm 2.9 \mu\text{mol L}^{-1}$, respectively), and there was no significant difference among depths in all sites (**Table 4.3**).

DO concentrations in the flooded burnt sites decreased between depths of 10 and 20 cm and became stable below 20 cm, suggesting the presence of DO supply from the ground surface (**Fig. 4.8**). Liebner *et al.* (2012) also observed a rapid decrease (from $>80\%$ to 0% of air saturation) in DO in the top 20 cm of soil in an alpine wetland and a constant DO profile below the top layer. Under tropical climate, Ueda *et al.* (2000) measured DO concentrations in the groundwater of a coastal peat swamp in Thailand, finding that most concentrations were below the detection limit and had a maximum value of 14 $\mu\text{mol L}^{-1}$. Given these observations, it is likely that DO in saturated peat soils remains very low even near the ground surface, similar to DO concentrations observed in this study (**Fig. 4.8**). Most of DO in peat water would be consumed in the top 10 cm, thus there was no significant differences among depths

at each study site. Previous studies also pointed out that the oxic zone only extends from the soil surface to water table or at most a few centimeters below (King *et al.*, 1990; Benstead and Lloyd, 1994; Nedwell and Watson, 1995; Whalen *et al.*, 1996).

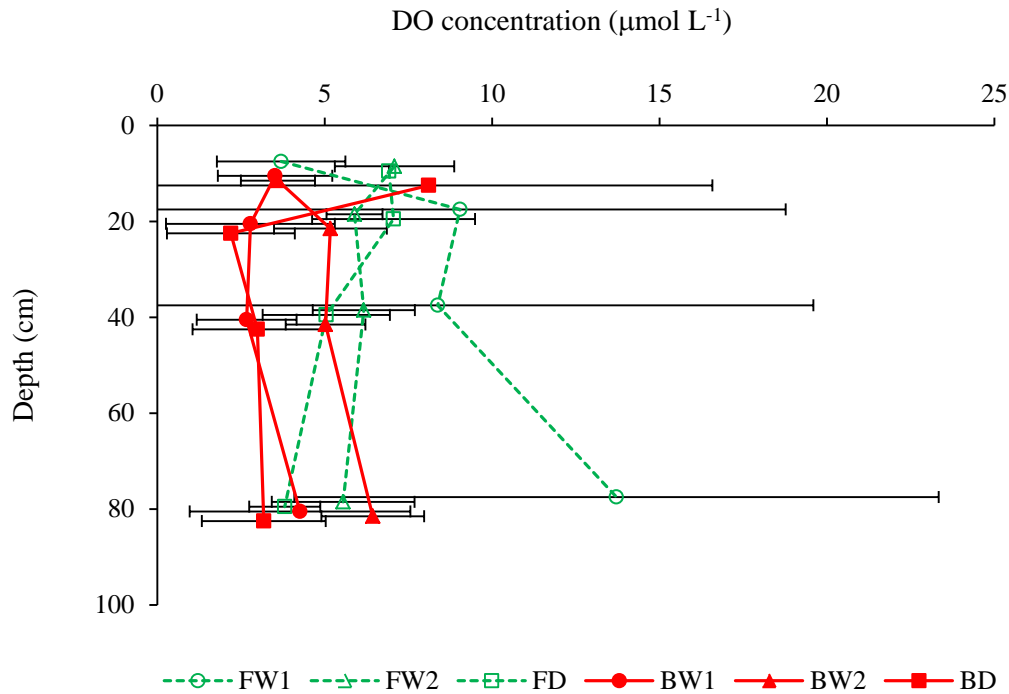


Fig. 4.8 Vertical distributions of DO concentration in subsurface pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation. For better identifiability, the depths in the profiles were slightly shifted from their actual depths (10, 20, 40, and 80 cm).

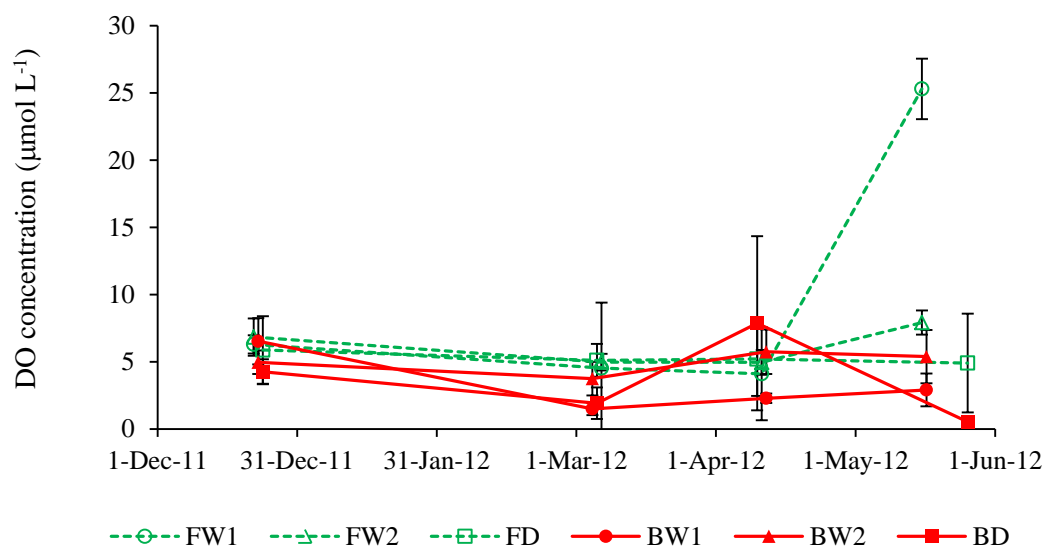


Fig. 4.9 Seasonal variations of DO concentration in subsurface pore water averaged for depths of 10–80 cm at each site. Error bars show standard deviation.

Table 4.2 DO concentration under different land use and drainage conditions averaged for the observation period.

Drainage	Depth	Land use					
		Natural forest		Burnt area		All	
Flooded condition	10 cm	5.63 ± 2.47	(7)	3.56 ± 1.33	(8)	4.53 ± 2.16	(15)
	20 cm	7.47 ± 6.61	(8)	3.98 ± 2.36	(8)	5.72 ± 5.12	(16)
	40 cm	7.27 ± 7.51	(8)	3.85 ± 1.78	(8)	5.56 ± 5.56	(16)
	80 cm	9.63 ± 7.79	(8)	5.19 ± 2.75	(7)	7.56 ± 6.23	(15)
	All	7.56 ± 6.38	(31)	4.11 ± 2.08	(31)	5.84 ± 5.02	(62)
Drained condition	10 cm	6.92 ± 0.00	(1)	8.11 ± 8.47	(3)	7.81 ± 6.94	(4)
	20 cm	7.05 ± 2.43	(3)	2.20 ± 1.91	(4)	4.28 ± 3.25	(7)
	40 cm	5.05 ± 1.90	(4)	2.99 ± 1.94	(4)	4.02 ± 2.09	(8)
	80 cm	3.81 ± 1.06	(4)	3.18 ± 1.85	(4)	3.49 ± 1.44	(8)
	All	5.29 ± 2.07	(12)	3.85 ± 4.19	(15)	4.49 ± 3.44	(27)
All	10 cm	5.79 ± 2.33	(8)	4.80 ± 4.49	(11)	5.22 ± 3.68	(19)
	20 cm	7.35 ± 5.64	(11)	3.38 ± 2.30	(12)	5.28 ± 4.61	(23)
	40 cm	6.53 ± 6.17	(12)	3.56 ± 1.79	(12)	5.05 ± 4.69	(24)
	80 cm	7.69 ± 6.87	(12)	4.46 ± 2.57	(11)	6.15 ± 5.41	(23)
	All	6.93 ± 5.59	(43)	4.03 ± 2.89	(46)	5.43 ± 4.62	(89)

1. Values are means and standard deviations of DO concentration in subsurface water ($\mu\text{mol L}^{-1}$).
2. Numbers in parentheses represent sample size at each level.

Table 4.3 Results of the three-way ANOVA on DO concentrations among different land use, drainage, and depth.

	DO	
	<i>df</i>	<i>F</i>
<i>Main factors</i>		
Land use	1	4.87 *
Drainage	1	0.66
Depth	3	0.20
Error	73	
<i>Interactions</i>		
Land use*Drainage	1	0.62
Land use*Depth	3	0.38
Drainage*Depth	3	1.31
Land use*Drainage*Depth	3	0.31

Symbols * and ** mean $P < 0.05$ and $P < 0.01$, respectively.

4.3.4 pH and EC

Peat water pH was relatively stable and ranged from 3.5 to 4.0 among the study sites. It was highest at BD followed by BW1>BW2>FW1>FW2>FD (Table 4.4; Fig. 4.10). There was almost no difference in pH among the depths.

The measured values of pH reflected the acidic nature of tropical peatlands. SOM contains a high amount of carboxyl groups and phenolic hydroxyl, which are important in exchange reactions and as sources of acidity (Thomas and Hargrove, 1984). During the breakdown and humification of organic materials, the insoluble and soluble products containing substituted radicals such as $-\text{COOH}$, phenol-OH, alcohol-OH, $-\text{NH}_2$ and quinines are released. The acidity developed as a consequence of the dissociation of hydrogen from the acidic groups of carboxyl and phenolic hydroxyl (Schnitzer and Gupta, 1965).

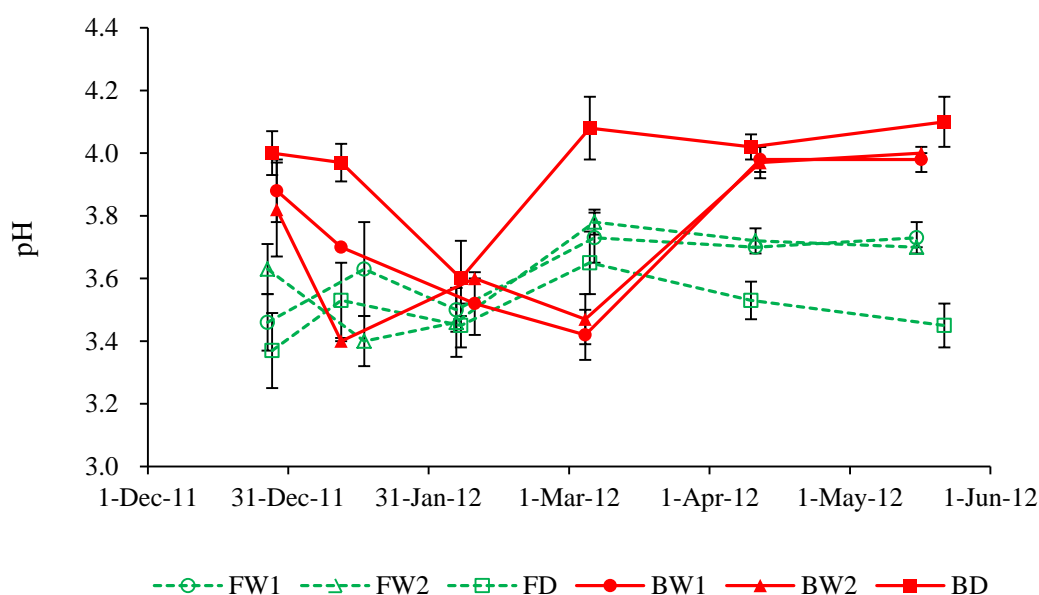


Fig. 4.10 Seasonal variations of pH in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation.

Peat water EC was also stable and ranged from 35 to 125 $\mu\text{S cm}^{-1}$ among the study sites and depths (Fig. 4.11). The EC was highest at FD, followed by FW1>FW2>BW1>BW2>BD (Table 4.4). Temporal change and difference among depths were also small relative to the difference among the sites.

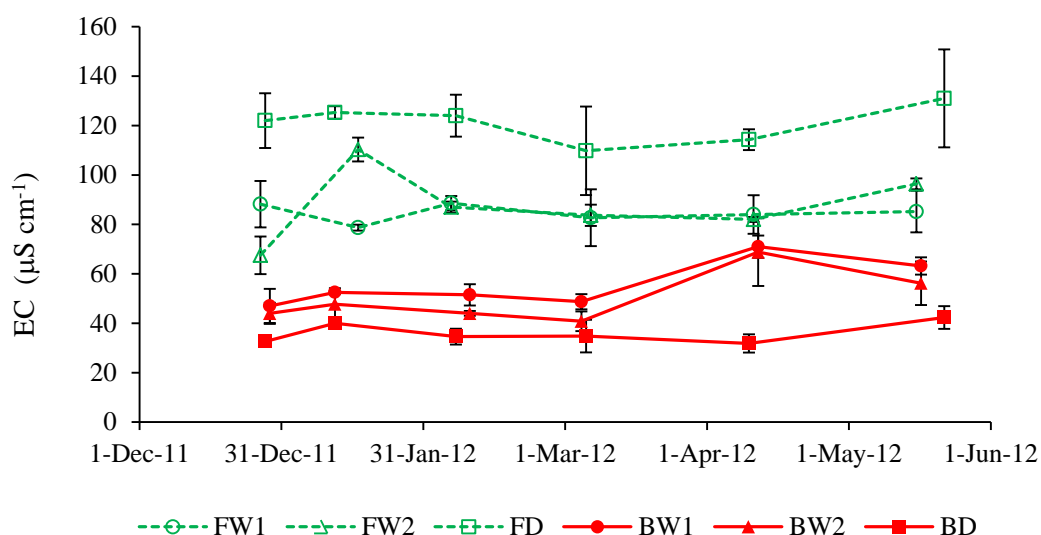


Fig. 4.11 Seasonal variations of EC in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation.

Table 4.4 pH and EC at all study sites averaged for the observation period (Dec. 2011 to May 2012).

Site	Description	pH	EC ($\mu\text{S cm}^{-1}$)
FW1	Flooded forest site #1	3.6 ± 0.1 ab (31)	85 ± 8 c (31)
FW2	Flooded forest site #2	3.6 ± 0.2 ab (31)	86 ± 14 c (31)
FD	Drained forest site	3.5 ± 0.1 a (17)	125 ± 16 d (17)
BW1	Flooded burnt site #1	3.7 ± 0.2 b (31)	56 ± 10 b (31)
BW2	Flooded burnt site #2	3.7 ± 0.2 b (32)	50 ± 12 b (32)
BD	Drained burnt site	4.0 ± 0.2 c (28)	35 ± 5 a (28)

1. Values are means and standard deviations of environmental factors.
2. Numbers in parentheses represent sample size at each site.
3. Values within the same column with different lowercase letters differ significantly among sites (Bonferroni, $P < 0.05$). The values increase in alphabetical order.

4.3.5 Dissolved nitrogenous ions

NO_3^- -N concentration ranged from 0.0519 to 0.260 mg L^{-1} . The mean of NO_3^- -N concentration was generally higher in the burnt sites than in the forest sites. NO_3^- -N was highest at BW1, followed by BW2 > BD > FD > FW1 > FW2 (**Table 4.5**). There was significant difference only between BW2 and the flooded forest sites (FW1 and FW2). At BW1 and BW2, NO_3^- -N concentration increased in March and decreased in May, while at other sites it was stable (**Fig. 4.12**).

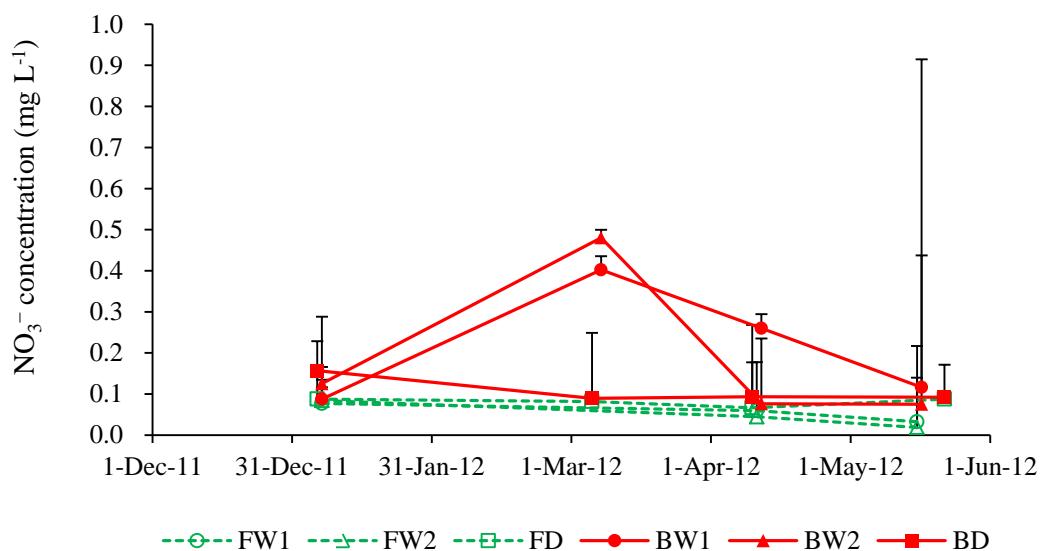


Fig. 4.12 Seasonal variations of NO_3^- concentration in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation (positive side only).

NH_4^+ -N concentration ranged from 0.061 to 0.205 mg L^{-1} . According to **Fig. 4.13**, there were small seasonal variations in NH_4^+ concentrations during field observation (Dec. 2011 to May 2012). In the late of May, the concentrations of NH_4^+ in peat water rose at all study sites. Sites BW1 and BW2 showed a rapid increase in NH_4^+ concentrations compared to other sites. There are no significant differences in NH_4^+ -N concentration among all sites (**Table 4.5**).

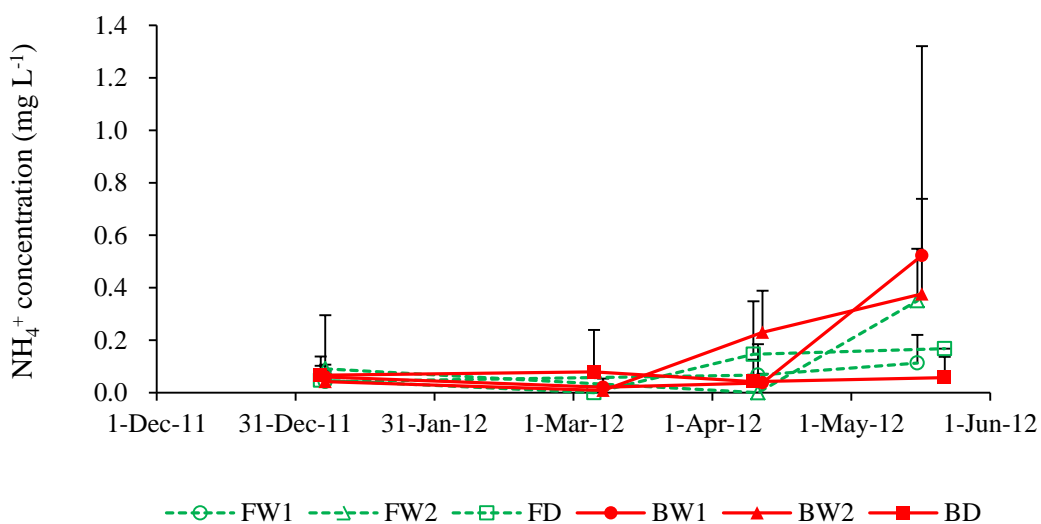


Fig. 4.13 Seasonal variations of NH_4^+ concentration in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation (positive side only).

Table 4.5 NO_3^- and NH_4^+ concentrations at all study sites averaged for the observation period (Dec. 2011 to May 2012).

Site	Description	NO_3^- -N (mg L^{-1})	NH_4^+ -N (mg L^{-1})
FW1	Flooded forest site #1	0.0565 ± 0.0492 a (15)	0.0741 ± 0.0928 (15)
FW2	Flooded forest site #2	0.0519 ± 0.0319 a (13)	0.143 ± 0.214 (13)
FD	Drained forest site	0.0781 ± 0.0325 ab (9)	0.086 ± 0.130 (9)
BW1	Flooded burnt site #1	0.260 ± 0.253 b (15)	0.194 ± 0.491 (15)
BW2	Flooded burnt site #2	0.211 ± 0.219 ab (15)	0.205 ± 0.263 (15)
BD	Drained burnt site	0.0915 ± 0.0401 ab (13)	0.061 ± 0.110 (13)

1. Values are means and standard deviations of environmental factors.
2. Numbers in parentheses represent sample size at each site.
3. Values within the same column with different lowercase letters differ significantly among sites (Bonferroni, $P < 0.05$). The values increase in alphabetical order.

In the forest sites, NH_4^+ -N concentrations were slightly higher than NO_3^- -N concentrations, and vice versa in the burnt sites (**Table 4.5**). This was in correspondent with peat water pH, which was generally lower in the forest sites than in the burnt sites (**Table 4.4**). In general, acidic soils tend to have more NH_4^+ than NO_3^- (Pearson *et al.*, 2002).

NH_4^+ and NO_3^- are the key substrates for the soil microbial processes nitrification and denitrification responsible for production and emission of N_2O . Thus the effects of NH_4^+ and NO_3^- are discussed in relation to the flux and dissolved concentration of N_2O in **Chapters 5 and 6**.

4.4 Summary

The investigations of environmental characteristics showed that: 1) the precipitation and water table level showed temporal changes during the observation period, 2) DO concentrations were occasionally changed at all sites, only the main effect of land use was significant ($P < 0.05$; forest > burnt, 6.9 ± 5.6 and $4.0 \pm 2.9 \mu\text{mol L}^{-1}$, respectively), and there was no significant difference among depths in all sites, 3) pH and EC peat water were relatively stable among the study sites. The pH in drained burnt site was highest and EC in drained forest site was highest, and 4) in the forest sites, NH_4^+ -N concentrations were slightly higher than NO_3^- -N concentrations, and vice versa in the burnt sites.

DYNAMICS OF GHG FLUXES UNDER DIFFERENT LAND USE AND WATER TABLE LEVELS

5.1 Introduction

Recently, large areas of tropical peatland have been developed for large-scale agricultural in Southeast Asia. Land use change and the drainage conditions will impact on the production of GHG in peatland. This chapter describes the impact of land use change and water table level conditions on the dynamics of GHG fluxes and their contributions on the GWP.

5.2 Materials and methods

5.2.1 Measurement of GHG fluxes

GHG fluxes at the ground surface were measured monthly using a closed-chamber method (Morishita *et al.*, 2003; Takakai *et al.*, 2006; Toma and Hatano, 2007). A set of cylindrical stainless steel chambers (18.5–21 cm in diameter and 25 cm in height), each with an acrylic lid fixed to a sample collector and a pressure-regulating bag, was used for the flux measurements. Stainless steel collars were permanently planted at three locations at each site.

First, 400 mL of the air inside the chamber was taken into a Tedlar[®] bag as the sample at 0 min before closing the chamber lid. The chamber was then quickly capped with the lid. After 6 min, 250 mL of the air inside the chamber was transferred into another Tedlar[®] bag. CO₂ concentrations in these bags were determined on the day of sampling with a non-dispersive infrared (NDIR) analyzer (ZFP9; Fuji Electric, Japan). The lid was then removed temporarily, and the air inside the chamber was ventilated. Next, 20 mL of the air inside the chamber was periodically collected into pre-evacuated vial bottles with butyl rubber stoppers at 0, 20, and 40 min after closing the lid for CH₄ and N₂O analyses. CH₄ concentrations were determined using a gas chromatographer (GC-8A, Shimadzu, Kyoto, Japan) equipped with 1) a flame ionization detector (FID) maintained at 130°C and 2) a 2-m-long activated carbon column (80/100 mesh; Stainless column, Serial No. D-0015, Shimadzu, Kyoto, Japan) maintained at 70°C with pure N₂ gas as a carrier. N₂O concentrations were determined with a gas chromatographer (GC-14B, Shimadzu, Kyoto, Japan) equipped with 1) an electron capture detector (ECD) maintained at 340°C and 2) a 1-m-long Porapak N column maintained at 60°C with proportional gas (5% CH₄ in Ar).

The gas flux was calculated on the basis of the change in gas concentration inside the chamber (ΔC_c) against closure time (Δt):

$$F = \rho_g \times (V / A) \times (\Delta C_c / \Delta t) \times [273 / (273 + T)] \quad (5.1)$$

where F is the gas flux (gC or gN m⁻² h⁻¹), ρ_g is the gas density (0.536×10^3 g m⁻³ for CO₂-C and CH₄-C; 1.259×10^3 g m⁻³ for N₂O-N), V/A is equivalent to the height of the chamber from the water or ground surface (m), and T is the air temperature (triplicated air temperature measurements near the chamber at 100 cm high) during the sampling period (°C). Air temperature was measured before closing and after opening the chamber, and the mean of both temperatures was used in **Eq. 5.1**. A positive flux indicates gas emissions from the soil or water surface into the atmosphere, while a negative flux indicates gas uptake from the atmosphere.

GHG fluxes were not measured when water table levels exceeded the height of the chamber. As a result, GHG fluxes at several sites were not measured in December 2011 (FW1, BW1, and BW2), February 2012 (BW1), and March 2012 (FW1, FW2, and BW1). Due to a separate technical issue, GHG fluxes could not be measured in January 2012 for any site, with the exception of FW2.

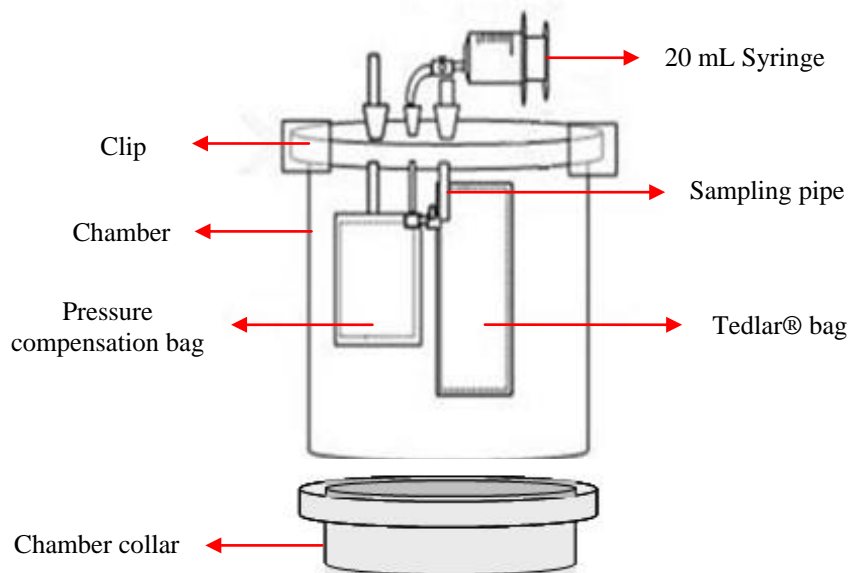


Fig. 5.1 Schematic diagram of a closed chamber (Toma and Hatano, 2007).

5.2.2 GWP calculation

The total flux of global warming potential (GWP) was calculated as CO₂-equivalent carbon flux (gCO₂-eq C m⁻² h⁻¹) by the following equation:

$$\text{GWP} = F_{\text{CO}_2} + F_{\text{CH}_4} \times (16/12) \times 25 \times (12/44) + F_{\text{N}_2\text{O}} \times (44/28) \times 298 \times (12/44) \quad (5.2)$$

where 25 and 298 are the factors for converting CH₄ and N₂O fluxes to their GWPs in a time horizon of 100 years (IPCC, 2007); (16/12), (44/28), and (12/44) are the conversion factors of C to CH₄, N to N₂O, and CO₂ to C, respectively.

5.2.3 Statistical analysis

The effects of land use (forest or burnt area) and drainage (flooded or drained) on GHG fluxes were tested by two-way ANOVA. Multiple comparison tests among the sites were conducted using the Bonferroni method. A probability level less than 0.05 ($P < 0.05$) was considered statistically significant. All analyses were performed in IBM SPSS Statistics Ver. 21.

5.3 Results

Temporal changes in the fluxes of CH₄, CO₂, and N₂O are shown in **Figs. 5.2, 5.4, and 5.6**, respectively. The relationships between water table levels and these fluxes are shown in **Figs. 5.3, 5.5, and 5.7**, respectively. These GHG fluxes were averaged for each group (classified by land use and drainage) and are summarized in **Tables 5.1–5.3**, with the results of two-way ANOVA in **Table 5.4**.

5.3.1 Soil CH₄ flux

According to **Table 5.1**, CH₄ fluxes were highest in the flooded burnt sites (5.75 ± 6.66 mgC m⁻² h⁻¹) followed by the flooded forest sites (1.37 ± 2.03 mgC m⁻² h⁻¹), the drained burnt site (0.220 ± 0.143 mgC m⁻² h⁻¹), and the drained forest site (0.0084 ± 0.0321 mgC m⁻² h⁻¹). The site-specific averages of CH₄ fluxes ranged from 0.0084 to 5.75 mgC m⁻² h⁻¹ (**Table 5.1**). The high CH₄ emissions at BW1 and BW2 and negative CH₄ fluxes (= CH₄ absorption) at the forest sites were periodically observed (**Fig. 5.2**). At FD site, 7 of 15 measured fluxes were negative. However, on average, all sites were a net CH₄ sources during the observation period (**Table 5.1**). In the burnt sites, higher average CH₄ fluxes were associated with the higher average water table levels (**Fig. 5.3**). Relative to the burnt sites, the CH₄ fluxes at the forest sites were consistently lower regardless of water table levels. The results of two-way ANOVA indicated that both main effects (land use and drainage) and their interaction were significant, and CH₄ flux at the flooded burnt sites was significantly higher than those of the other sites (**Table 5.4**). This means the CH₄ flux under flooded conditions was significantly higher than under drained conditions in both land use during the field observation.

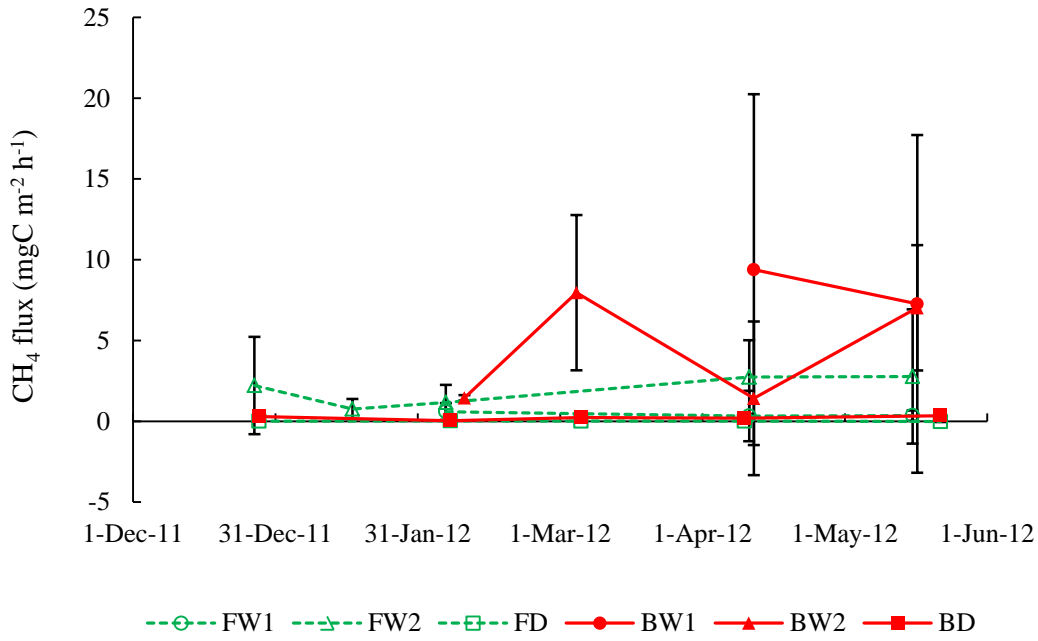


Fig. 5.2 Seasonal variations of CH₄ fluxes measure at the ground surface in the study sites. Error bars show standard deviation.

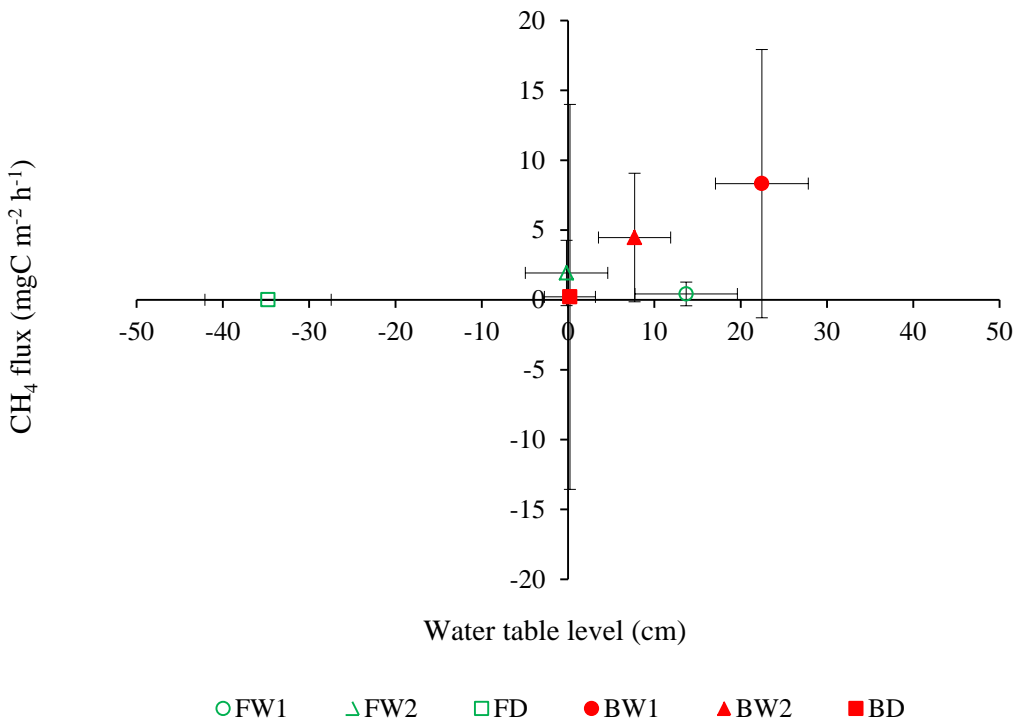


Fig. 5.3 Relationship between CH₄ flux and water table level averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

Table 5.1 CH₄ flux measured at the ground surface under different land use and drainage conditions averaged for the observation period.

Drainage	Land use		
	Natural forest	Burnt area	All
Flooded	1.37 ± 2.03 (24)	5.75 ± 6.66 (18)	3.25 ± 5.05 (42)
Drained	0.0084 ± 0.0321 (15)	0.220 ± 0.143 (15)	0.114 ± 0.148 (30)
All	0.84 ± 1.72 (39)	3.24 ± 5.60 (33)	1.94 ± 4.14 (72)

1. Values are means and standard deviations of CH₄ flux (mgC m⁻² h⁻¹) at the ground surface. Positive values mean net emission.
2. Numbers in parentheses represent sample size at each level.

5.3.2 Soil CO₂ flux

The site-specific averages of CO₂ fluxes ranged from 108 to 340 mgCm⁻² h⁻¹ (**Table 5.2**). The CO₂ flux in the drained forest site was highest (340±250 mgC m⁻² h⁻¹ with the water table level of -20 to -60 cm), which was followed by those in the flooded burnt sites and flooded forest sites (198±165 mgC m⁻² h⁻¹ with -5 to +40 cm and 195±199 mgC m⁻² h⁻¹ with -10 to +30 cm, respectively), and was significantly higher than that in the drained burnt site (108±115 mgC m⁻² h⁻¹ with -15 to +10 cm). CO₂ fluxes tended to be relatively low (<150 mgC m⁻² h⁻¹) and stable until Feb. 2012, with higher values (>350 mgC m⁻² h⁻¹) occurring after Mar. 2012 at FW2, FD and BD (**Fig. 5.4**). Among the study sites, CO₂ flux was highest at FD, except for in Apr. 2012. In the forest sites, slightly lower average CO₂ fluxes were associated with higher average water table levels (**Fig. 5.5**). In contrast, the largest CO₂ flux was associated with the highest water table level at BW1. However, this trend may have been an artifact of sampling effort: the CO₂ flux was not measured at BW1 from Dec. 2011 to Feb. 2012, when this flux was generally low at all other sites. According to the results of two-way ANOVA, the effect of land use and the interaction of land use*drainage were significant, and CO₂ flux at the drained forest site was significantly higher than that at the drained burnt site (**Table 5.4**).

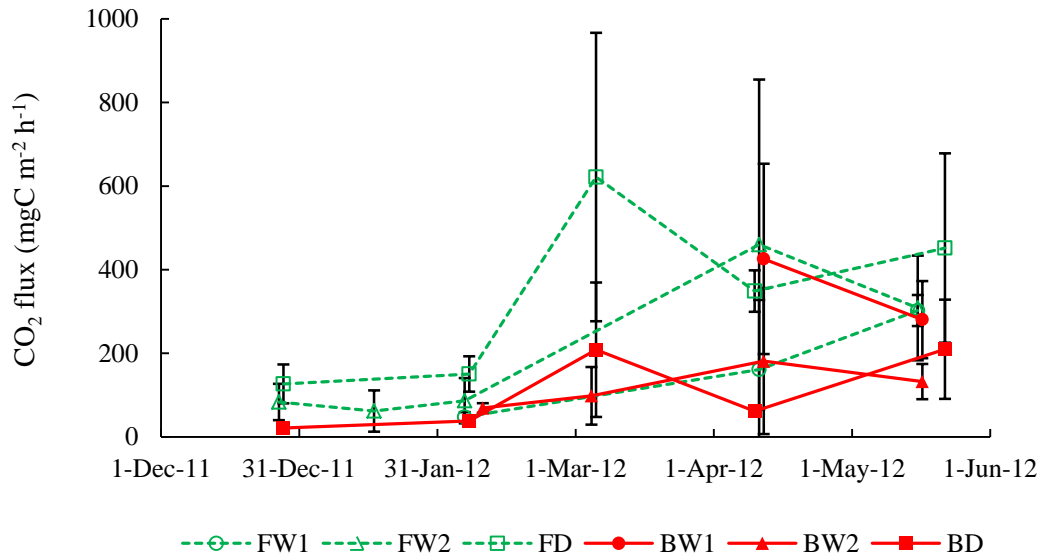


Fig. 5.4 Seasonal variations of CO₂ fluxes measure at the ground surface in the study sites. Error bars show standard deviation.

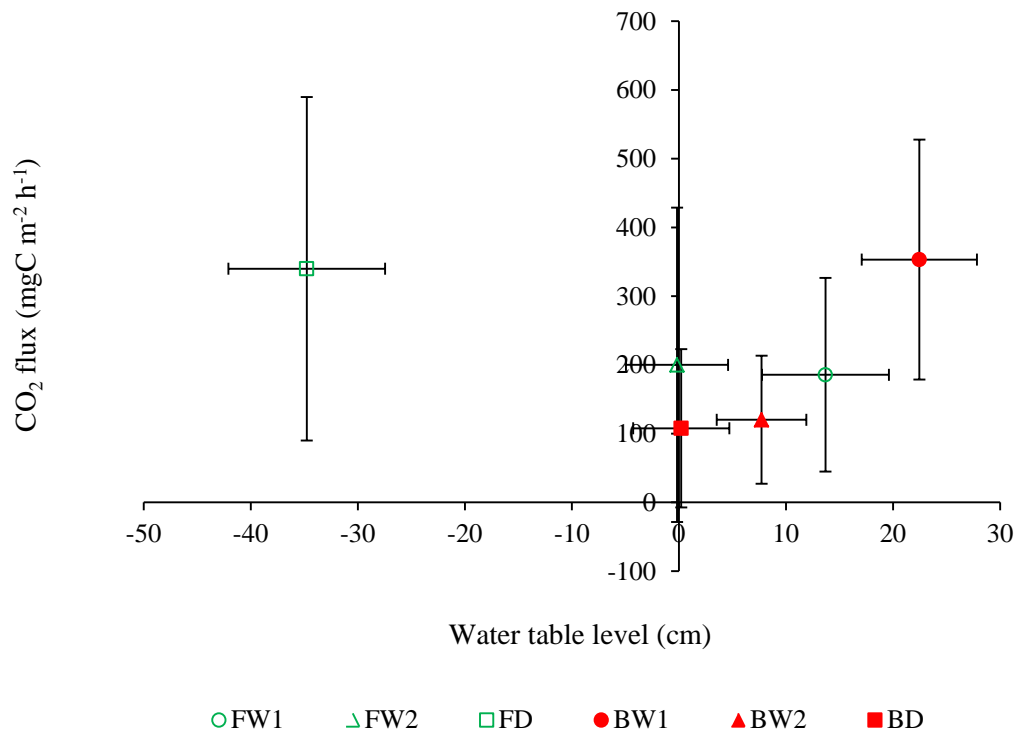


Fig. 5.5 Relationship between CO₂ flux and water table level averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

Table 5.2 CO₂ flux measured at the ground surface under different land use and drainage conditions averaged for the observation period.

Drainage	Land use		
	Natural forest	Burnt area	All
Flooded	195 ± 199 (23)	198 ± 165 (18)	196 ± 183 (41)
Drained	340 ± 250 (15)	108 ± 115 (15)	224 ± 225 (30)
All	252 ± 229 (38)	157 ± 150 (33)	208 ± 200 (71)

1. Values are means and standard deviations of CO₂ flux (mgC m⁻² h⁻¹) at the ground surface. Positive values mean net emission.
2. Numbers in parentheses represent sample size at each level.

5.3.3 Soil N₂O flux

The site-specific averages of N₂O flux ranged from -8.7 ± 41.9 to 8.1 ± 75.5 $\mu\text{gN m}^{-2} \text{h}^{-1}$ (**Table 5.3**). Although occasional variations were observed in early Apr. at BD and BW2, little variation was observed in N₂O flux at other sites during the observation period (**Fig. 5.6**). On average, the N₂O fluxes at the drained sites tended to be positive, whereas they tended to be negative at flooded sites. This difference suggested that drainage influenced whether a site acted as a sink or a source of N₂O in tropical peatlands. However, the relationship between N₂O flux and water table level was unclear (**Fig. 5.7**). The results of two-way ANOVA suggested that there were no significant differences in N₂O fluxes between sites based on either land use or drainage (**Table 5.4**). The N₂O flux measured at each chamber at study sites ranged from -155 to 218 $\mu\text{gN m}^{-2} \text{h}^{-1}$.

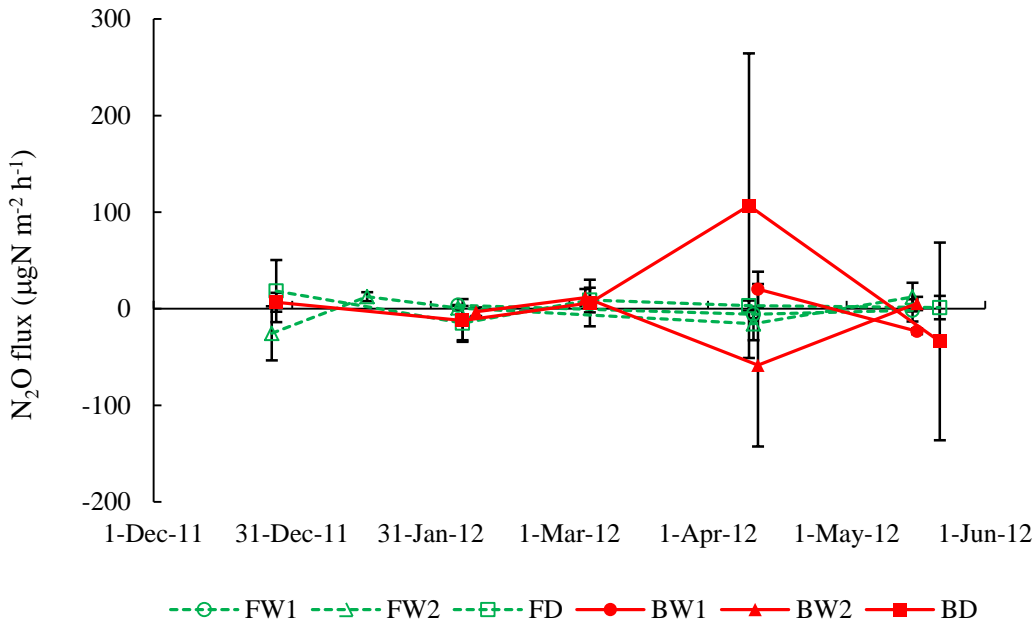


Fig. 5.6 Seasonal variations of N₂O fluxes measure at the ground surface in the study sites. Error bars show standard deviation.

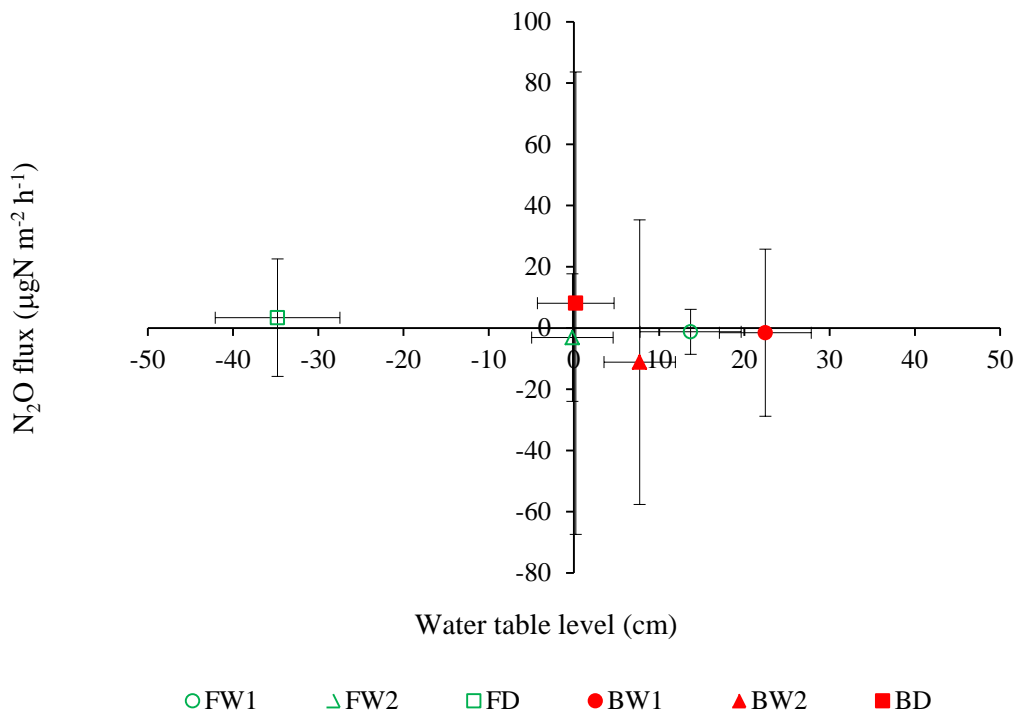


Fig. 5.7 Relationship between N₂O flux and water table level averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

Table 5.3 N₂O flux measured at the ground surface under different land use and drainage conditions averaged for the observation period.

Drainage	Land use		
	Natural forest	Burnt area	All
Flooded	-2.4 ± 16.9 (24)	-8.7 ± 41.9 (16)	-4.9 ± 29.2 (40)
Drained	3.4 ± 19.2 (15)	8.1 ± 75.5 (14)	5.7 ± 53.3 (29)
All	-0.2 ± 17.8 (39)	-0.9 ± 59.5 (30)	-0.5 ± 41.0 (69)

1. Values are means and standard deviations of N₂O flux ($\mu\text{gN m}^{-2} \text{h}^{-1}$) at the ground surface. Positive values mean net emission.
2. Numbers in parentheses represent sample size at each level.

Table 5.4 Summary of ANOVA on GHG fluxes and GWP under different land use and drainage conditions and multiple comparison among the study sites.

Sites	CH ₄ flux (mgC m ⁻² h ⁻¹)	CO ₂ flux (mgC m ⁻² h ⁻¹)	N ₂ O flux (µgN m ⁻² h ⁻¹)	GWP (CO ₂ -eq mgC m ⁻² h ⁻¹)			
Flooded forest sites (FW1 and FW2)	1.37 ± 2.03 a (24)	195 ± 199 ab (23)	-2.4 ± 16.9 (24)	208 ± 204 ab (23)			
Drained forest site (FD)	0.0084 ± 0.0321 a (15)	340 ± 250 b (15)	3.4 ± 19.2 (15)	340 ± 250 b (15)			
Flooded burnt sites (BW1 and BW2)	5.75 ± 6.66 b (18)	198 ± 165 ab (18)	-8.7 ± 41.9 (16)	249 ± 171 ab (18)			
Drained burnt site (BD)	0.220 ± 0.143 a (15)	108 ± 115 a (15)	8.1 ± 75.5 (14)	111 ± 113 a (15)			
ANOVA	df	F	df	F	df	F	
<i>Main factors and interaction</i>							
Land use	1	7.34 **	1	6.33 *	1	0.01	4.13 *
Drainage	1	16.49 **	1	0.36	1	1.23	0.00
Land use*Drainage	1	6.05 *	1	6.64 *	1	0.29	8.59 **
Error	68		67		67		
<i>Simple main effects</i>							
Land use under flooded condition		15.85 **		0.00		0.22	0.47
Land use under drained condition		0.03		11.30 **		0.09	10.73 **
Drainage in forest vegetation		1.36		5.32 *		0.18	4.34 *
Drainage in burnt area		20.05 **		1.86		1.23	4.26 *

1. Fluxes are shown in means±standard deviations. Positive fluxes mean net emission from peat soil.
2. Numbers in parentheses represent sample sizes in each level.
3. Values within the same column with different lowercase letters differ significantly ($P < 0.05$; corrected by Bonferroni method).
4. Symbols * and ** mean $P < 0.05$ and $P < 0.01$, respectively.

5.3.4 GWP

GWP was mostly affected by CO₂ flux, followed by CH₄ flux. The GWP equivalent to CO₂ flux ranged from 110.84 to 428.69 mgC m⁻² h⁻¹ over the measuring period from Dec. 2011 to May 2012 (**Fig. 5.8**). At BW1 and BW2, 18 to 25% of the GWP were attributed to CH₄, while most of the GWP (>92%) was due to CO₂ at other sites. As a result, averaged GWP at the flooded burnt sites was 20% larger compared to the flooded forest sites, although the CO₂ fluxes at both sites were similar (**Table 5.5**). This exception was probably due to the fact that CH₄ emissions in the flooded burnt sites were larger than those in the drained burnt site. The results of two-way ANOVA were similar to those for the analysis of CO₂ flux, except that the simple main effect of drainage in the burnt area was significant (**Table 5.4**).

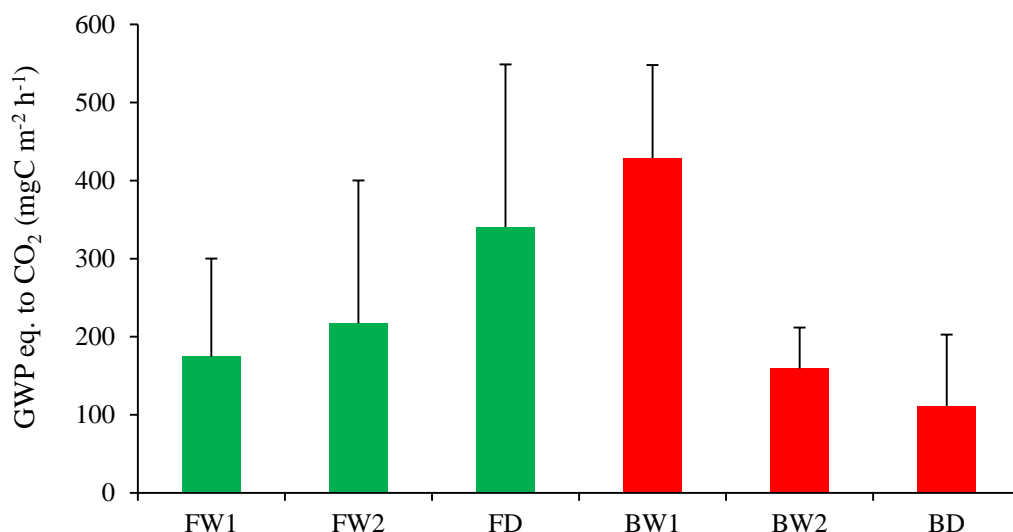


Fig. 5.8 GWP equivalent to CO₂-C flux at the ground surface averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

Table 5.5 GWP eq. to CO₂-C fluxes measured at the ground surface under different land use and drainage conditions averaged for the observation period.

Drainage	Land use		
	Natural forest	Burnt area	All
Flooded	208 ± 204 (23)	249 ± 171 (18)	226 ± 189 (41)
Drained	340 ± 250 (15)	111 ± 113 (15)	226 ± 224 (30)
All	260 ± 230 (38)	186 ± 161 (33)	226 ± 203 (71)

1. Values are means and standard deviations of GWP as CO₂ equivalent carbon flux (mgC m⁻² h⁻¹) at the ground surface. Positive values mean net emission.
2. Numbers in parentheses represent sample size at each level.

5.4 Discussion

5.4.1 Dynamics of soil CH₄ flux in tropical peatlands

5.4.1.1 Effect of water table level

In this study, especially in burnt sites, CH₄ fluxes were mainly influenced by drainage conditions (**Table 5.4 and Fig. 5.3**), consistent with the results of previous studies (Martikainen *et al.*, 1992; Moore and Dalva, 1993). Positive relationships between mean water table level and CH₄ emissions have been reported in non-flooded northern peatlands (e.g., Bubier and Moore, 1993; Pelletier *et al.*, 2007). When water table is present above the ground surface, it impedes entry of O₂ to support methanotrophy (Bartlett *et al.*, 1988; Koschorreck, 2000).

Based on a comprehensive review on CH₄ flux observation in tropical peatlands, Couwenberg *et al.* (2010) has recently reported that CH₄ fluxes are generally low and often distinctly negative for water levels below -20 cm, while tend to be higher and more variable at higher water levels. Jauhiainen *et al.* (2005) found that CH₄ fluxes were positive at water table levels >-50 cm and were negative at levels <-60 cm at locations that were in the vicinity of our study area. Jauhiainen *et al.* (2008) also reported that CH₄ fluxes began to increase as the water table levels rose higher than -40 to -20 cm in a drained forest or higher than -30 to 0 cm in a deforested, burned site. Melling *et al.* (2005a) showed that CH₄ fluxes ranged from -4.53 to 8.40 μgC m⁻² h⁻¹ in a mixed peat swamp forest (water table level: -60 to -20 cm), from -7.44 to 102 μgC m⁻² h⁻¹ in a sago plantation (-40 to 0 cm), and from -32.8 to 4.17 μgC m⁻² h⁻¹ in an oil palm plantation (-80 to -30 cm) and also increased with increasing water table levels.

In drained forest site, in contrast, CH₄ flux was low probably due to the lower water table levels, which was always kept at -20 to -60 cm. This drained condition might lower the CH₄ production in the drained forest site. Under such aerobic conditions, methanotrophic activity would be promoted instead of methanogenesis. The relationship between ground water level and methane emissions is comparable to the analogy of an “on-off switch” given by Christensen *et al.* (2003). The deeper the water table level was, the more the peat soil becomes susceptible to aerobic decomposition. Methanogenesis, which is a strictly anaerobic process, may also decrease following the increases in O₂ during drainage. The magnitude of this decrease depends on the depth of drainage and thus the depth of oxic zone in soil profile (Furukawa *et al.*, 2005; Martikainen *et al.*, 1995).

CH₄ fluxes and trends in the drained sites were similar to those reported in these previous studies (**Table 5.4**). Considering the relationship between CH₄ flux and water table level cited above, there is an empirical depth of water table level at which CH₄ production and oxidations are balanced. In this study, the drained forest site was both a weak sink and source of CH₄ (**Table 5.1 and Fig. 5.3**) with a water table level that ranged from -39.7 to -7.0 cm at the time of flux measurement, similar to the water levels observed in the previous studies.

5.4.1.2 Impact of land use change

The role of plants in the production and transport of CH₄ from tropical peatlands is complex. Aerenchyma promoted gas exchange between the atmosphere and the rhizosphere (Megonigal and Day, 1992; Kozłowski, 1997), in particular entry of O₂ to the root zone. The rhizosphere is often a site of major microbiological activity (Curl and Truelove, 1986) and subsurface emission of O₂ can have profound effects on the balance of organic degradation that occurs in waterlogged soils (Nedwell *et al.*, 1994). In recent years, biotic factors such as vegetation are considered to control CH₄ emissions from wetlands, because two gas-transport mechanism, molecular diffusion and convective gas flow, are well documented in some plants (Brix *et al.*, 1992; Nouchi and Mariko, 1993). In addition, near-surface O₂ level is affected by transport linked to the presence of plants and O₂ release from plant roots (Elberling *et al.*, 2011).

CH₄ fluxes were lower in the forest sites relative to the burnt sites under both flooded and drained conditions (**Table 5.1**). Meanwhile DO concentrations were influenced by land use only and were significantly higher in the forest sites compared to the burnt sites (**Table 4.2 and 4.3**). These results suggested that CH₄ produced in the peat might be oxidized by plant-mediated oxygen supply in the forest sites. Blodau and Moore (2003) observed a depletion of O₂ and initiation of CH₄ production after flooding unsaturated peat in a bog mesocosm. Methanogenesis are potentially influenced by the effects of DO concentrations in soil profiles.

In the burnt sites, the water table level always near or above the surface during the wet season even in the drainage-affected site (**Fig. 4.5**). This was partly due to land subsidence after deforestation and drainage, which had resulted in the lack of new C input, loss of surface peat by aerobic decomposition and burning, and shrinkage and compaction of peat layer. Depression patches, which had depths of 20–30 cm and was created formerly by smoldering fire in the burnt sites, had remained unvegetated possibly due to surface wetness

during wet seasons (Jauhiainen *et al.*, 2008). Thus, land use change not only directly affects the GHG dynamics and C balance, but also may indirectly influence them by changing relative heights between the ground surface and water level.

5.4.2 Dynamics of soil CO₂ flux in tropical peatlands

In tropical peatlands, Melling *et al.* (2005b) reported that soil CO₂ fluxes ranged from 100 to 533 mgC m⁻² h⁻¹ in a mixed peat swamp forest, from 63 to 245 mgC m⁻² h⁻¹ in a sago plantation, and from 46 to 335 mgC m⁻² h⁻¹ in an oil palm plantation in Sarawak, Malaysia. Jauhiainen *et al.* (2005) also reported that CO₂ fluxes were 132 to 166 mgC m⁻² h⁻¹ in hummocks and 37.9 to 188 mgC m⁻² h⁻¹ in hollows in a tropical peat swamp forest in Central Kalimantan. The CO₂ fluxes obtained in this study were comparable to those found in previous studies.

5.4.2.1 Effect of water table level

Drainage and development of tropical peatland leads to drastic hydrological changes and CO₂ emissions to atmosphere (Hooijer *et al.*, 2006; Page *et al.*, 2002). Local hydrology is known to strongly influence peat decomposition (e.g. Hirano *et al.*, 2009; Page *et al.*, 2009). Because organic carbon sequestration in wetlands is dependent on slow decomposition rates, drainage and the associated increases in O₂ diffusion into wetlands sediments often increase ecosystem respiration, leading to a net oxidation of organic material (Bridgham and Richardson, 1992; Freeman *et al.*, 1993a; b; Furukawa *et al.*, 2005; Hirano *et al.*, 2007; Jauhiainen *et al.*, 2008; Kluge *et al.*, 2008; Martikainen *et al.*, 1995) and an increase in CO₂ emission (Moore and Knowles, 1989; Bubier, 1995; Silvola *et al.*, 1996; Nykänen *et al.*, 1998).

This study showed the effect of lowered water table level on the emissions of CO₂, which were most typically found in the drained forest site. CO₂ fluxes were highest in the drained forest site followed by the flooded burnt sites, flooded forest sites and drained burnt site (**Table 5.2 and 5.4**). This is because the drained forest site has a larger oxic zone due to drainage. According to the data of water table level, it was always kept -20 to -60 cm at the drained forest site (**Fig. 4.5**). Drainage to greater depths increases the area of profile oxidation and, thus, tends to increase C loss (Armentano and Menges, 1986; Furukawa *et al.*, 2005). Jali (2004) and Hadi *et al.* (2005) have noted the oxic conditions after drainage are known to promote microbial decomposition in peat.

In the flooded forest sites, in contrast, the inhibition of aerobic peat decomposition would suppress CO₂ emissions. In the flooded burnt sites, a continuously flooded condition would inhibit CO₂ diffusion into the atmosphere and cause the high dissolved CO₂ concentrations at 20 cm. In contrast, CH₄ production at depth of 20 cm was higher in the flooded burnt sites (**Table 6.1 and 6.2**). The ratio of CO₂/CH₄ in saturated soil was similar, when the CO₂ are consumed by methanogens. This was caused by decreasing of respirations to oxidized H₂ due to low DO concentrations. At that condition, hydrogenotrophic methanogens use CO₂ to produce CH₄ in the soil by used H₂ as an electron donor. The CO₂ fluxes in the drained burnt site, which were significantly lower than those in the drained forest site, suggested a low root respiration rate from poor vegetation cover. Insufficient drainage relative to the drained forest site may have also inhibited CO₂ production through peat decomposition in the drained burnt site.

The increasing trends in CO₂ flux during the observation period, which differed from the trends in CH₄, N₂O, and other environmental factors including water table level, could be attributed to the increase in plant root respiration as opposed to microbial peat decomposition. The acclimation of plant roots to saturated conditions may be one possible explanation (Drew *et al.*, 1994; Mommer *et al.*, 2004). Because roots may acclimate metabolically to a gradual fall in O₂ supply, so that they either improve their tolerance of anoxia, or partially avoid O₂-deficiency by structural changes that aid internal transfer of O₂ to the roots from the shoot (Drew, 1990).

5.4.2.2 Impact of land use change

In this study, the drained forest site was significantly higher contribution in CO₂ emission (**Table 5.2**). The differences in CO₂ flux between different land uses could probably be due to their belowground biomass and productivity. Soil CO₂ emission, as the results of soil respiration is generated mainly from autotrophic (root) and heterotrophic (microbial) activity (Janssens *et al.*, 2001). With bigger biomass and productivity, we could expect larger root mass, and therefore a higher root respiration can be expected (Silvola *et al.*, 1996). The amount of CO₂ produced by roots would be greater when there are more living roots in the soil. Thus, the higher CO₂ emission in the drained forest site would reflect the direct effect of larger root biomass, as well as its lowest water table level.

Since roots are also the major source of additional C input to the soil, the direct effect of elevated root respiration can be accompanied by enhanced soil microbial respiration in the immediate vicinity of the roots, the rhizosphere. Roots produce highly decomposable fine roots and root exudates that can be rapidly decomposed by microbes when exposed to more aerobic conditions (Thomas *et al.*, 1996). According to Lohila *et al.* (2003) there would be an increase in the amount of soil microorganism utilizing the root exudates in the rhizosphere, thus enhancing the soil respiration activities. Based on these previous studies, higher CO₂ emission under dense vegetation of peat swamp forest may partly be attributed to this indirect effect of plant roots.

5.4.3 Dynamics of soil N₂O flux in tropical peatlands

Like the fluxes in CH₄, the N₂O fluxes observed in this study were generally related to drainage condition. Melling *et al.* (2007) reported that the N₂O flux ranged from -3.4 to 19.7 µgN m⁻² h⁻¹ in a mixed swamp forest, from 1.0 to 176.3 µgN m⁻² h⁻¹ in a sago plantation, and from 0.9 to 58.4 µgN m⁻² h⁻¹ in an oil palm plantation. The N₂O fluxes in the mixed peat swamp forest in that study were comparable to those observed in the drained sites of this study. Takakai *et al.* (2006) also measured N₂O flux at locations that were identical several of those observed in this study. According to their results, the average N₂O fluxes in the wet season (2002 to 2004) were 49±63 µgN m⁻² h⁻¹ in the drained forest site and 55±100 µgN m⁻² h⁻¹ in the drained burnt site, comparable to the observations made in the wet season (2011–2012) in this study.

5.4.3.1 Effect of water table level

The low N₂O emissions at all study sites could be influenced by water table level. Although the results of statistical analysis showed that there was no significant different in N₂O fluxes both in land use change and drainage conditions, it was found that the drained sites in this study acted as N₂O sources, while the flooded sites acted as N₂O sink. This phenomenon might occur due to the flooded conditions which reduce the gaseous exchange between soil and atmosphere and enhance the complete denitrification of N₂O to N₂ (Granli and Boeckman, 1994; Inubushi *et al.*, 2003; Mosier and Delgado, 1997; Sehy *et al.*, 2003), resulting in lower emission of N₂O to the atmosphere (Denmead *et al.*, 1979). Therefore as suggested by Denmead *et al.* (1979), flooded soils may contribute less N₂O but more N₂ to the atmosphere. In this study, the flooded burnt sites consumed more N₂O than the flooded

forest sites (**Table 5.3**). This could be attributed to differences in DO concentrations, with significantly higher concentrations in the forest sites compared to the burnt sites (**Table 4.3; Fig. 4.8**).

This absorption of atmospheric N₂O into peat water for denitrification is also related to the amount of NO₃⁻ (Burgin and Groffman, 2012). A recent review indicated that net negative N₂O fluxes have been reported in numerous previous studies, showing that low mineral nitrogen and high moisture content are favorable for N₂O consumption (Chapuis-Lardy *et al.*, 2007). NO₃⁻ concentrations observed in this study (**Table 4.5**) were much lower than those observed in crop lands in the vicinity of the sites in this study (Takakai *et al.* 2006), supporting the possibility of net N₂O consumption in the flooded shallow peat.

5.4.3.2 Impact of land use change

Soil is considered to be one of the key sources of N₂O emitted into the atmosphere as it contributes approximately 60% of the emissions globally (Ehhalt *et al.*, 2001). The application of N fertilizer generally increases N₂O emission rate from the crop land. The increasing use of inorganic N fertilizers under humid tropical climate may contribute to N₂O production significantly (Hadi *et al.*, 2000; Takakai *et al.*, 2006). Land use change from pristine peat swamp forest to agricultural use such as cropland and plantation will therefore increase net N₂O emission.

NO₃⁻ concentration is one of the important controlling factors for soil N processes, including nitrification and denitrification. In this study, the effect of NO₃⁻ on N₂O flux was unclear, although there were significant differences in dissolved NO₃⁻ concentrations in peat water among the study sites (**Table 4.5**). Takakai *et al.* (2006) found that the N₂O flux in a cropland in the vicinity of our study area increased with increasing NO₃⁻-N content in the top 10 cm of soil during rainy season. In that study, however, the NO₃⁻-N content was significantly higher in cropland (200 to 300 mg kg⁻¹ dry soil) than in the drained forest or in the drained burnt sites (0.43 to 91 and 0.79 to 5.9 mg kg⁻¹ dry soil, respectively). In this study, the total NO₃⁻-N content was likely too small to influence N₂O emissions. In addition, Inubushi *et al.* (2003) have investigated seasonal changes in N₂O emission in N₂O emission over a whole year in tropical peatland in South Kalimantan. N₂O emission from abandoned agriculture land and secondary forest were low (ranging from -40 to 30 µgN m⁻² h⁻¹), and they found no clear seasonal changes in N₂O emission. They explained this result as inhibition of N₂O emission by flood water.

5.4.4 Effect of each GHG flux on GWP

In this study, GWP mostly affected by CO₂ flux, followed by CH₄ flux and N₂O flux. The highest CO₂ fluxes in drained forest site were affected by the high CO₂ production in peat due to lower water table level (**Fig. 4.5**). In addition, the water table level at this site was consistently at 20–60 cm below the ground surface, much lower than other sites. The lower water table increases organic matter availability for aerobic (CO₂-releasing) decomposition. Consequently, GWPs in the drained forest site were higher compared with other sites.

GWPs from the soil in the burnt sites were generally lower compared with those in the forest sites (**Fig. 5.8; Table 5.5**). Low DO concentration (**Table 4.2**), and probably less amount of plant roots and low substrate quality of the recalcitrant peat in the flooded burnt sites might inhibit microbial activity for aerobic peat decomposition and the following CO₂ emission. At the flooded burnt sites, however, CH₄ fluxes were higher due to the higher production of CH₄ under flooded conditions and the lower DO concentrations. As a result, in spite of almost the same CO₂ flux, GWP in the flooded burnt sites was 20% higher than that in the flooded forest sites due to the large CH₄ emission, although it was not significant. The low N₂O emissions at all study sites could be influenced by water table level. In drained sites, soils could play a role as a source, meanwhile in flooded sites could play a role as a sink of N₂O. The low N₂O flux made little contribution to GWP.

5.5 Summary

The investigations of field study showed that: 1) CH₄ fluxes were significantly influenced by land use and drainage, which were highest in the flooded burnt sites (5.75 ± 6.66 mgC m⁻² h⁻¹) followed by the flooded forest sites (1.37 ± 2.03 mgC m⁻² h⁻¹), the drained burnt site (0.220 ± 0.143 mgC m⁻² h⁻¹), and the drained forest site (0.0084 ± 0.0321 mgC m⁻² h⁻¹), 2) CO₂ fluxes were significantly higher in the drained forest site (340 ± 250 mgC m⁻² h⁻¹ with the water table level of -20 to -60 cm) than in the drained burnt site (108 ± 115 mgC m⁻² h⁻¹ with the water table level of -15 to +10 cm), and 3) N₂O fluxes ranged from -2.4 to -8.7 μgN m⁻² h⁻¹ in the flooded sites and from 3.4 to 8.1 μgN m⁻² h⁻¹ in the drained sites. The negative N₂O fluxes might be caused by N₂O consumption by denitrification under flooded conditions. GWP was mainly determined by CO₂ flux, with highest levels in the drained forest site.

DYNAMICS OF DISSOLVED GHG CONCENTRATIONS UNDER DIFFERENT LAND USE AND WATER TABLE LEVELS

6.1 Introduction

The land use change and drainage conditions were affected on the dynamics of dissolved GHG concentrations. In an effort to understand the relationship between the productions of these gases in the soil, this chapter describes the dynamics of dissolved GHG concentrations in the soil and their contribution on GHG emissions to the atmosphere.

6.2 Materials and methods

To determine the concentration of GHGs dissolved in peat water, two types of in-situ headspace method were applied: a rapid equilibration method for CH₄ and N₂O and in-situ equilibration chamber for CO₂.

6.2.1 Concentrations of dissolved CH₄ and N₂O

Gas samples for the analyses of dissolved CH₄ and N₂O were obtained by rapidly equilibrating a fixed volume of air with pore water in a syringe (**Fig. 6.1**). Water samples were collected from depths of 20, 40, 60, and 80 cm. Pore water was extracted from the same type of water sampler used for the measurement of DO and water quality (see **Chapter 4**). During the observation intervals, the sampler was filled with water, and the stopcock was kept closed to prevent the exchanges of GHGs and O₂ between the atmosphere and soils at these depths.

Before sampling, the water remaining in the sampler was removed. First, 30 mL was collected from the sampler directly into a 60-mL plastic syringe. Next, the same volume of O₂-free air was injected into the same syringe. The syringe was then shaken by hand for two minutes (100 times). After that, 20 mL of the air that was equilibrated with the water sample was stored in a pre-evacuated vial. The concentrations of CH₄ and N₂O in the bottles were determined in the same manner used for flux measurements for these gases.



Fig. 6.1 View of the rapid equilibration method for the measurement of dissolved CH_4 and N_2O concentrations.

6.2.2 Concentration of dissolved CO_2

To determine the concentrations of CO_2 dissolved in peat water, an in-situ equilibration chamber was applied. An in-situ equilibration chamber (i.e., a small chamber designed to equilibrate air with the surrounding dissolved gases) was developed according to previous studies (Faulkner *et al.*, 1989; Yu and DeLaune, 2006). An inverted 60-mL plastic syringe was used as the chamber and was connected aboveground to a thin tube, which was equipped with a three-way stopcock at the upper end. A chamber was installed at the same depths for CH_4 and N_2O measurement (i.e. 20, 40, 60, and 80 cm).

First, 60 mL of O_2 -free air was injected through the tube, and the stopcock was closed to keep the injected air from escaping. Two to four weeks following the injection, the air was extracted from the chamber and collected into an aluminum sample bag. CO_2 concentration was determined on the day of sampling by the same method applied for the CO_2 flux measurement described before (in **Chapter 5**). In some case CO_2 concentration exceeded the upper range of the NDIR analyzer (= 2000 ppmv). In that case the air sample was diluted with CO_2 -free gas to 1:10 or 1:100.



Fig. 6.2 View of field settings for dissolved CO₂ concentrations measurements at the study site.

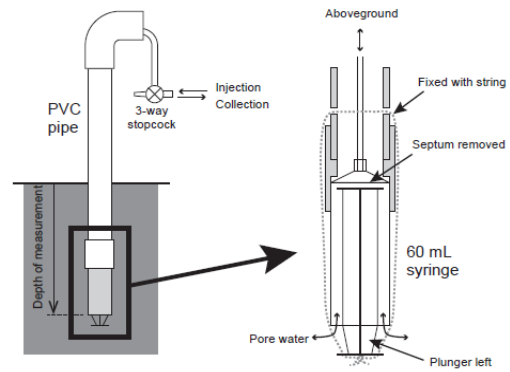


Fig. 6.3 Diagram of an in-situ equilibration chamber in the field.

The time required for CO₂ equilibrium was investigated beforehand at the same sites in this study as follows. The CO₂ concentration in the air samples collected following different equilibration durations (5 to 100 days) were compared. As a result, no difference in dissolved CO₂ was statistically significant among the durations. The initial O₂-free air was injected just after air collection in the last month. Therefore, CO₂ in the collected samples were all expected to be in equilibrium with CO₂ dissolved in the surrounding water.

6.2.3 Calculations of dissolved GHG concentrations

The dissolved gas concentration in pore water (C_l , mol mol⁻¹) in this method (dissolved CH₄, CO₂, and N₂O concentrations) was given by Henry's law:

$$C_l = \frac{C_g \times 10^{-6} \times P}{K_H} \quad (6.1)$$

where C_g is the concentration of the sample air (ppmv), P is the total pressure inside the chamber (atm), and K_H is Henry's constant (mol mol⁻¹ atm⁻¹). The effect of water pressure above the chamber was reflected in P . The value of K_H , a function of soil temperature, was cited from a chemical engineering handbook (Society of Chemical Engineers, Japan, 1999). Soil temperature was preliminarily observed at depths of 5, 10, 30 and 50 cm (see **Chapter 4**). The average temperatures between 10 and 30 cm and between 30 and 50 cm were used as the soil temperatures at 20 and 40 cm, respectively. Soil temperature at 60 and 80 cm was assumed to be the same as that at 50 cm. The errors of ±1.6°C in soil temperature produced

differences of $\pm 4.2\%$ in estimation of dissolved CO_2 by **Eq. 6.1**. The C_l in molar ratio was converted to C_l (mol L^{-1}) according to the following equation:

$$C_l [\text{mol L}^{-1}] = \frac{C_l [\text{mol mol}^{-1}] \times \rho_w}{m_w} \quad (6.2)$$

where ρ_w is the water density (1000 g L^{-1}), and m_w is the molecular weight of water (H_2O , 18.01 g mol^{-1}).

According to the rapid equilibration method, the dissolved gas concentrations (C_l) in pore water could be calculated using the following equation (Alberto *et al.* 2000):

$$C_l = \frac{C_g (V_g + \alpha \cdot V_l) - C_{bg} \cdot V_g}{V_l} \quad (6.3)$$

where C_{bg} is the initial concentration of the gas in the O_2 -free air, α is the water:air partition coefficient, and V_g and V_l are the volume of the gaseous and liquid phases, respectively.

Eq. 6.3 can be simplified provided that 1) $C_g \gg C_{bg}$, and 2) $V_g = V_l$:

$$C_l = C_g (1 + \alpha) \quad (6.4)$$

Here, α is defined as C'_l/C_g , where C'_l is the dissolved gas concentration in the liquid phase inside the syringe. The variable C'_l was calculated in the same manner as C_l in **Eqs. 6.1 and 6.2**, assuming that the temperature inside the syringe was equal to T , which was air temperature observed near the flux chamber (see **Chapter 5**). The variable C_g was also converted to the same unit of C'_l as follows:

$$C_g [\text{mol L}^{-1}] = \frac{C_g [\text{ppmv}] \times 10^{-6}}{V} \quad (6.5)$$

where V (L mol^{-1}) is the volume of 1-mol air at total pressure P ($= 1.0 \text{ atm}$) and temperature T ($^\circ\text{C}$), can be calculated as:

$$V = 22.4 \times \frac{1}{P} \times \frac{T + 273}{273} \quad (6.6)$$

6.2.4 Statistical analysis

The effects of land use, drainage, and depth on dissolved GHG concentrations were tested using three-way ANOVA. Multiple comparison tests among the sites and depths were conducted using the Bonferroni method. A probability level less than 0.05 ($P < 0.05$) was

considered statistically significant. All analyses were performed in IBM SPSS Statistics Ver. 21.

6.3 Results

Seasonal variations in the dissolved concentrations of CH₄, CO₂, and N₂O during the observation period are shown in **Figs. 6.4, 6.6, and 6.8**, respectively. Meanwhile the vertical distributions of dissolved concentrations of these GHGs are shown in **Figs. 6.5, 6.7, and 6.9**, respectively. Dissolved GHGs at each depth were averaged for each group (classified by land use and drainage) and summarized in **Tables 6.1–6.3**, with the results of three-way ANOVA in **Table 6.4**.

6.3.1 Dissolved CH₄ concentration

Dissolved CH₄ concentrations were highest in the flooded burnt sites (124±84 μmol L⁻¹) followed by the drained burnt site (45.2±29.8 μmol L⁻¹), the flooded forest sites (1.15±1.38 μmol L⁻¹), and the drained forest site (0.860±0.819 μmol L⁻¹) (**Table 6.1**). Dissolved CH₄ occasionally changed in the late of Jan. 2012 at BW1 and BW2 then it became stable at all sites in the end of Jan. 2012 to May 2012 (**Fig. 6.4**). After Jan. 2012, dissolved CH₄ was clearly higher at BW1 and BW2, followed by BD. On the average basis, this relationship was common among all depth (**Table 6.1 and Fig. 6.5**). The dissolved CH₄ concentrations in the flooded burnt sites were 1–4 orders of magnitude higher than those at the forest sites. Dissolved CH₄ at the forest sites generally increased with increasing depth. The drained burnt site also showed a similar trend from a depth of 20 to 60 cm, but the flooded burnt sites showed an inverse profile, in which the dissolved CH₄ decreased with depth.

The result of the statistical analysis (**Table 6.4**) shows that dissolved CH₄ concentration significantly affected by all main factors, namely land use, drainage, and depth ($P < 0.01$ in all cases), meanwhile all the two-way interactions were significant ($P < 0.01$ in all cases). The results of the post hoc test were as follows. The effect of land use was significant at all depths under flooded condition ($P < 0.001$, forest < burnt). Under the drained condition, dissolved CH₄ concentration at 80 cm was nearly significant ($P = 0.059$, forest < burnt). The effect of drainage was significant only in the burnt sites at 20, 40, and 80 cm ($P < 0.001$ in all cases; flooded > drained). The effect of depth was significant only in the drained burnt site ($P < 0.001$; 20 cm > others depths). In short, dissolved CH₄ concentrations in the flooded burnt sites was significantly higher than those in the other sites, especially much higher at 20 cm. Dissolved

CH₄ in the drained burnt sites also significantly higher than those in the forest sites except for 20 cm.

Table 6.1 Dissolved CH₄ concentrations under different land use and drainage conditions averaged for the observation period.

Drainage	Depth	Land use		
		Natural forest	Burnt area	All
Flooded condition	20 cm	0.579 ± 0.631 (14)	183 ± 111 b (14)	92 ± 121 (28)
	40 cm	0.991 ± 0.611 (14)	128 ± 62 a (13)	62.4 ± 77.6 (27)
	60 cm	1.51 ± 1.40 (14)	85.9 ± 53.4 a (14)	43.7 ± 56.7 (28)
	80 cm	1.51 ± 2.15 (14)	100 ± 67 a (14)	50.7 ± 68.5 (28)
	All	1.15 ± 1.38 (56)	124 ± 84 (55)	62.1 ± 85.3 (111)
Drained condition	20 cm	0.345 ± 0.362 (7)	8.24 ± 5.22 (7)	4.3 ± 5.4 (14)
	40 cm	0.478 ± 0.447 (7)	52.9 ± 24.0 (7)	26.7 ± 31.7 (14)
	60 cm	0.705 ± 0.507 (7)	71.6 ± 27.9 (7)	36.2 ± 41.4 (14)
	80 cm	1.91 ± 0.79 (7)	48.1 ± 11.0 (7)	25.0 ± 25.1 (14)
	All	0.860 ± 0.819 (28)	45.2 ± 29.8 (28)	23.0 ± 30.6 (56)
All	20 cm	0.501 ± 0.557 (21)	125 ± 123 (21)	63 ± 106 (42)
	40 cm	0.820 ± 0.603 (21)	102 ± 63 (20)	50.2 ± 67.3 (41)
	60 cm	1.24 ± 1.23 (21)	81.1 ± 46.2 (21)	41.2 ± 51.7 (42)
	80 cm	1.64 ± 1.80 (21)	82.6 ± 60.1 (21)	42.1 ± 58.7 (42)
	All	1.05 ± 1.22 (84)	97.5 ± 79.6 (83)	49.0 ± 74.0 (167)

1. Values are means and standard deviations of dissolved CH₄ concentration (μmol L⁻¹).
2. Numbers in parentheses represent sample size at each level.
3. Values in the same land use and drainage with different lowercase letters differ significantly among depths (Bonferroni, *P*<0.05). The values increase in alphabetical order.

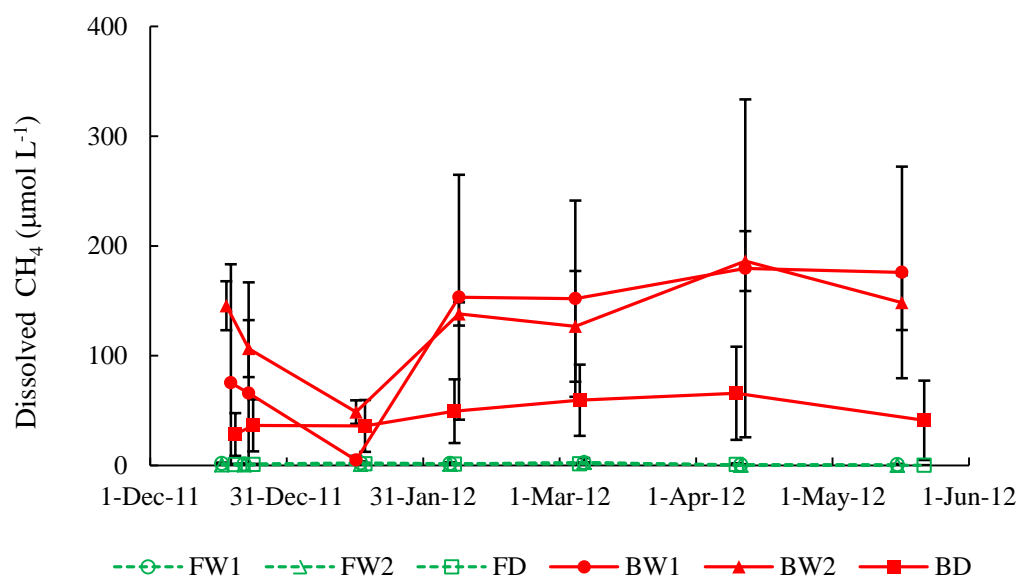


Fig. 6.4 Seasonal variations of dissolved CH₄ concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation.

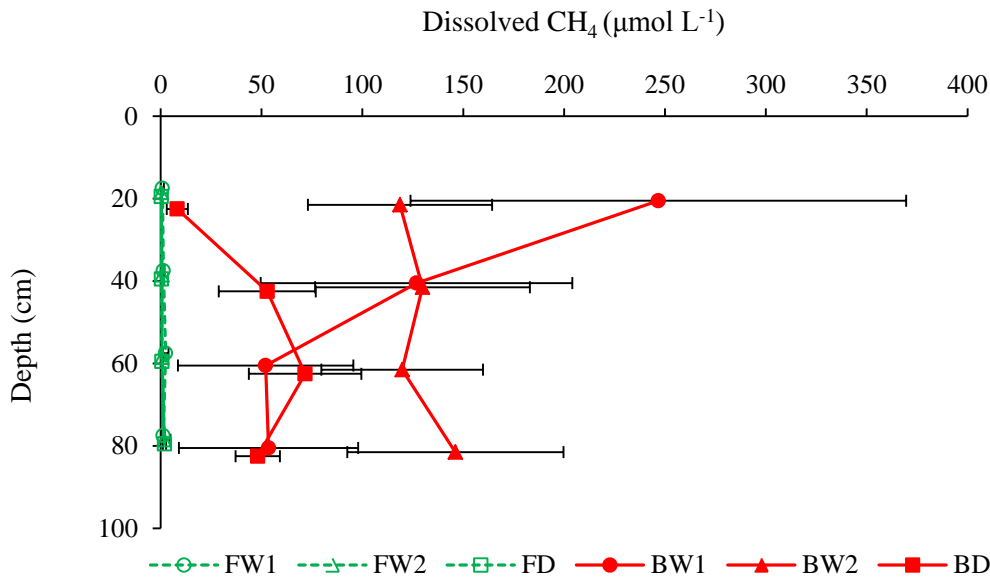


Fig. 6.5 Vertical distributions of dissolved CH₄ concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

6.3.2 Dissolved CO₂ concentration

Seasonal variations in dissolved CO₂ concentration during the observation period (Dec. 2011 to May 2012) ranged from 0.280 to 3.853 mmol L⁻¹. Dissolved CO₂ gradually increased at all sites, especially after Mar. 2012 (up to about 4 mmol L⁻¹ at FD; **Fig. 6.6**). This trend was similar to the seasonal change in CO₂ flux (**Fig. 5.4**). FD showed an increasing trend of dissolved CO₂ concentration with depth, and highest at depths of 40, 60, and 80 cm ($P < 0.01$, $P < 0.01$, and $P < 0.001$, respectively). In contrast, BW1 showed a decreasing trend of dissolved CO₂ concentration with depth, and highest at a depth of 20 cm ($P < 0.001$). Other sites did not show clear vertical trends.

In the calculation for dissolved CO₂ concentrations (**Eq. 6.1**), the daily mean soil temperature averaged for each depth and site was applied (**Fig. 4.7**). However, the largest diurnal variation in soil temperature at a depth of 20 cm was 3.2°C, recorded at FD in July 2011. This means that the difference between the daily average and the actual hourly soil temperature was as much as ±1.6°C. The errors of ±1.6°C in soil temperature produced differences of ±4.2% in estimation of dissolved CO₂ by **Eq. 6.1**. Because diurnal variations in soil temperatures were less than 3.2°C for all other sites and at the lower depths, errors associated with the use of daily mean soil temperatures in calculations were less than ±4.2% in the case of this study.

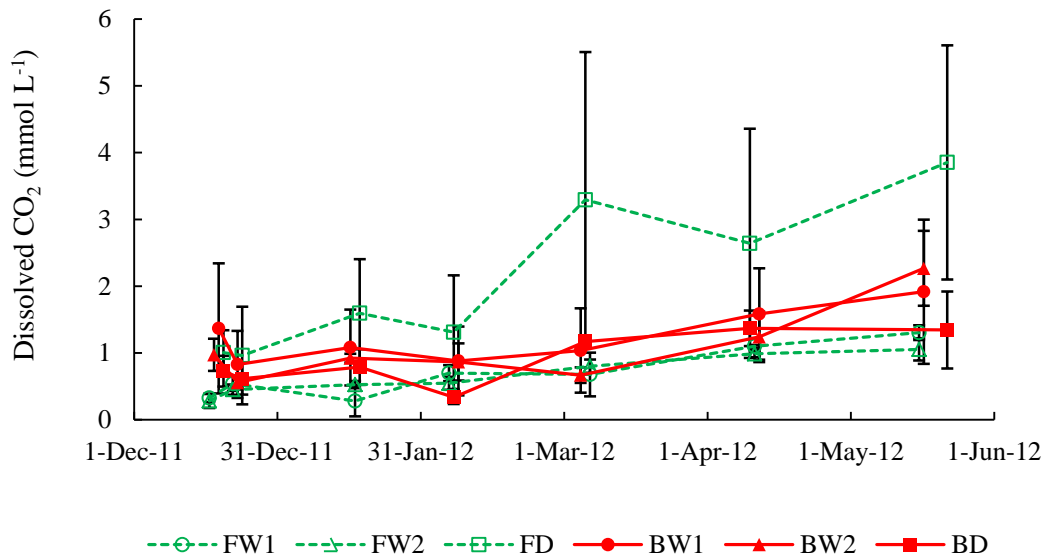


Fig. 6.6 Seasonal variations of dissolved CO₂ concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation.

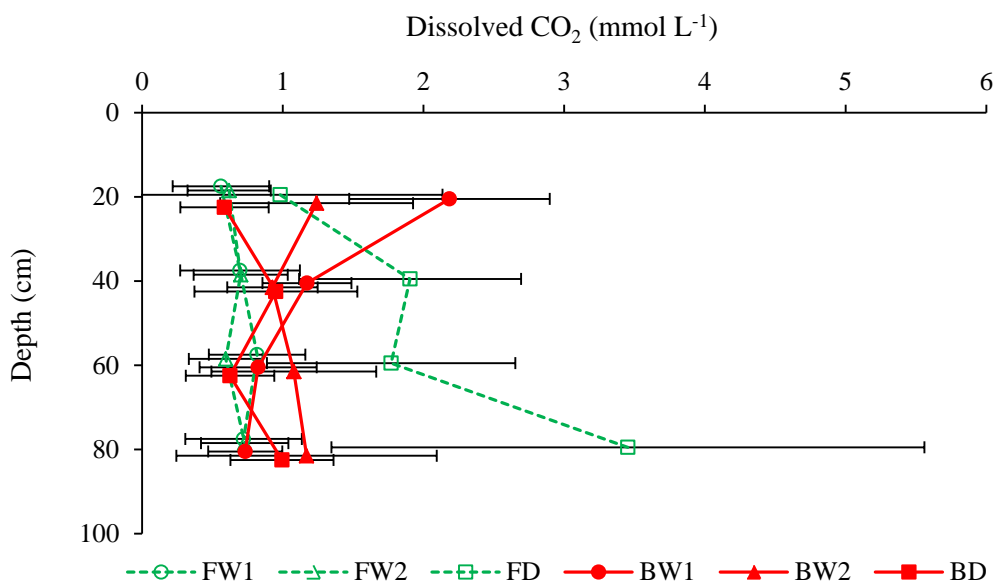


Fig. 6.7 Vertical distributions of dissolved CO₂ concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

The result of the three-way ANOVA (**Table 6.4**) shows that the effects of land use ($P < 0.01$), drainage ($P < 0.01$) and depth ($P < 0.01$) were significant. At the same time, all two-way interactions were also significant ($P < 0.01$ in all cases).

Table 6.2 Dissolved CO₂ concentrations under different land use and drainage conditions averaged for the observation period.

Drainage	Depth	Land use				
		Natural forest		Burnt area		All
Flooded condition	20 cm	0.591 ± 0.306	(13)	1.80 ± 0.829	(12)	1.17 ± 0.861 (25)
	40 cm	0.698 ± 0.368	(14)	1.06 ± 0.332	(12)	0.864 ± 0.390 (26)
	60 cm	0.698 ± 0.309	(13)	0.957 ± 0.507	(14)	0.832 ± 0.436 (27)
	80 cm	0.724 ± 0.354	(13)	0.939 ± 0.667	(13)	0.831 ± 0.534 (26)
	All	0.678 ± 0.330	(53)	1.17 ± 0.689	(51)	0.922 ± 0.589 (104)
Drained condition	20 cm	0.950 ± 1.11	(7)	0.597 ± 0.321	(6)	0.787 ± 0.835 (13)
	40 cm	1.88 ± 0.779	(7)	0.972 ± 0.591	(7)	1.43 ± 0.816 (14)
	60 cm	1.78 ± 0.887	(6)	0.640 ± 0.322	(3)	1.40 ± 0.918 (9)
	80 cm	3.47 ± 2.12	(7)	1.02 ± 0.375	(6)	2.34 ± 1.98 (13)
	All	2.03 ± 1.58	(27)	0.836 ± 0.453	(22)	1.49 ± 1.344 (49)
All	20 cm	0.717 ± 0.694	(20)	1.40 ± 0.903	(18)	1.04 ± 0.861 (38)
	40 cm	1.09 ± 0.773	(21)	1.03 ± 0.431	(19)	1.06 ± 0.627 (40)
	60 cm	1.04 ± 0.741	(19)	0.901 ± 0.487	(17)	0.974 ± 0.629 (36)
	80 cm	1.69 ± 1.82	(20)	0.963 ± 0.581	(19)	1.33 ± 1.39 (39)
	All	1.13 ± 1.14	(80)	1.07 ± 0.643	(73)	1.10 ± 0.937 (153)

1. Values are means and standard deviations of dissolved CO₂ concentration (mmol L⁻¹).
2. Numbers in parentheses represent sample size at each level.
3. Values in the same land use and drainage with different lowercase letters differ significantly among depths (Bonferroni, $P < 0.05$). The values increase in alphabetical order.

6.3.3 Dissolved N₂O concentration

The ranged of dissolved N₂O concentration was 0.000134 to 0.552 μmol L⁻¹, and it was occasionally changed at several sites (**Fig. 6.8**). Most of dissolved N₂O was <0.01 μmol L⁻¹ (142 data/167 in total). Only five data of dissolved N₂O were >0.1 μmol L⁻¹: four of them were obtained at depths of 20 cm and 40 cm of FD site in Jan.2012 (0.763 and 1.37 μmol L⁻¹) and Mar.2012 (0.226 and 0.104 μmol L⁻¹).

The result of three-way ANOVA (**Table 6.4**) showed that the effects of land use ($P < 0.05$) and drainage ($P < 0.05$) and the interaction of land use*drainage ($P < 0.05$) were significant. The effect of land use was significant at depths of 20 and 40 cm under drained condition ($P < 0.05$ and $P < 0.01$, respectively; forest>burnt), while the effect of drainage was significant at the same depths in the forest sites ($P < 0.05$ and $P < 0.01$, respectively; flooded<drained). Namely, dissolved N₂O concentration in the drained forest site was significantly higher than other sites at 20 and 40 cm (**Table 6.3**). There was no significant difference among depths at all sites (**Table 6.4**).

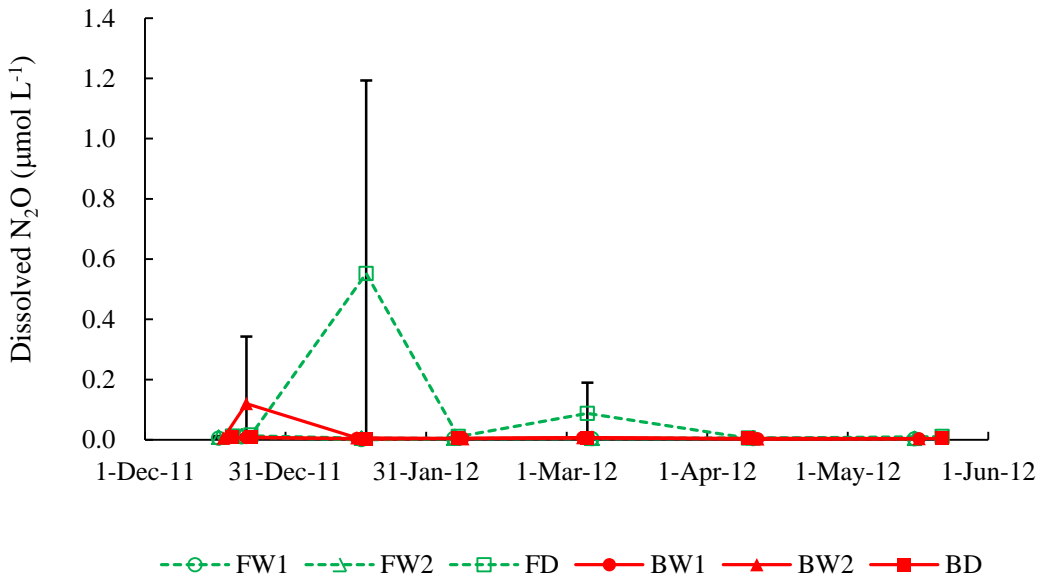


Fig. 6.8 Seasonal variations of dissolved N_2O concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation.

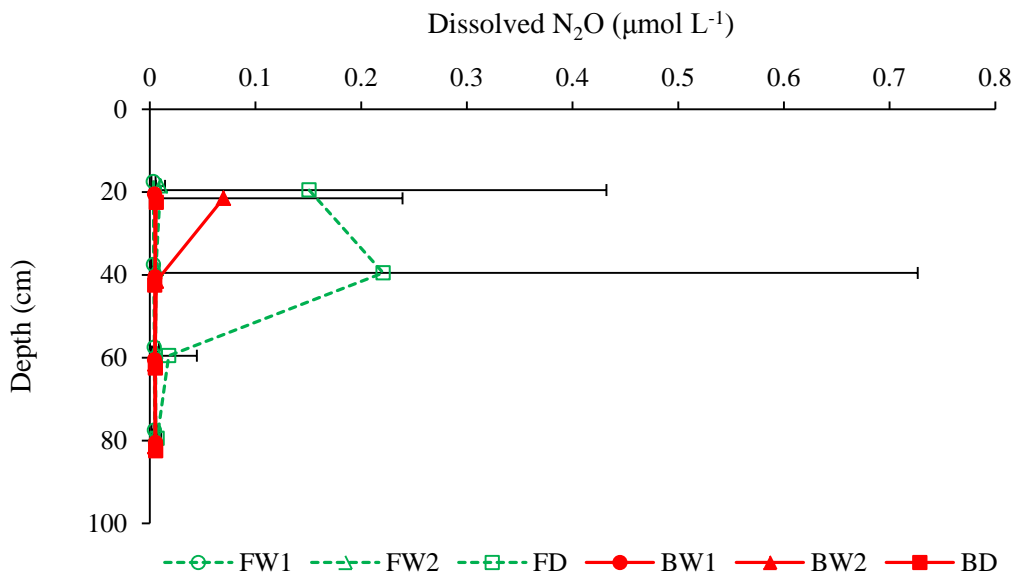


Fig. 6.9 Vertical distributions of dissolved N_2O concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

Table 6.3 Dissolved N₂O concentrations under different land use and drainage conditions averaged for the observation period.

Drainage	Depth	Land use		
		Natural forest	Burnt area	All
Flooded condition	20 cm	0.007 ± 0.005 (14)	0.037 ± 0.120 (14)	0.022 ± 0.085 (28)
	40 cm	0.005 ± 0.003 (14)	0.006 ± 0.003 (13)	0.005 ± 0.003 (27)
	60 cm	0.005 ± 0.004 (14)	0.005 ± 0.002 (14)	0.005 ± 0.003 (28)
	80 cm	0.005 ± 0.003 (14)	0.005 ± 0.001 (14)	0.005 ± 0.002 (28)
	All	0.005 ± 0.004 (56)	0.013 ± 0.061 (55)	0.009 ± 0.043 (111)
Drained condition	20 cm	0.15 ± 0.28 (7)	0.006 ± 0.002 (7)	0.079 ± 0.205 (14)
	40 cm	0.22 ± 0.51 (7)	0.005 ± 0.002 (7)	0.113 ± 0.361 (14)
	60 cm	0.018 ± 0.027 (7)	0.005 ± 0.003 (7)	0.012 ± 0.019 (14)
	80 cm	0.007 ± 0.004 (7)	0.006 ± 0.003 (7)	0.006 ± 0.003 (14)
	All	0.099 ± 0.288 (28)	0.006 ± 0.002 (28)	0.052 ± 0.207 (56)
All	20 cm	0.055 ± 0.169 (21)	0.027 ± 0.098 (21)	0.041 ± 0.137 (42)
	40 cm	0.077 ± 0.296 (21)	0.005 ± 0.003 (20)	0.042 ± 0.212 (41)
	60 cm	0.009 ± 0.016 (21)	0.005 ± 0.002 (21)	0.007 ± 0.012 (42)
	80 cm	0.006 ± 0.003 (21)	0.005 ± 0.002 (21)	0.006 ± 0.003 (42)
	All	0.037 ± 0.170 (84)	0.011 ± 0.049 (83)	0.024 ± 0.126 (167)

1. Values are means and standard deviations of dissolved N₂O concentration (μmol L⁻¹).
2. Numbers in parentheses represent sample size at each level.

Table 6.4 Results of three-way ANOVA on the dissolved GHG concentrations among different land use, drainage, and depth.

	Dissolved CO ₂		Dissolved CH ₄		Dissolved N ₂ O	
	df	F	df	F	df	F
<i>Main factors</i>						
Land use	1	8.54 **	1	125.95 **	1	4.70 *
Drainage	1	14.48 **	1	28.25 **	1	4.73 *
Depth	3	4.09 **	3	0.41	3	1.96
Error	137		151		151	
<i>Interactions</i>						
Land use*Drainage	1	47.98 **	1	27.84 **	1	6.56 *
Land use*Depth	3	6.70 **	3	0.50	3	1.59
Drainage*Depth	3	9.36 **	3	5.26 **	3	1.57
Land use*Drainage*Depth	3	1.67	3	5.30 **	3	1.95

Symbols * and ** mean $P < 0.05$ and $P < 0.01$, respectively.

6.4 Discussion

6.4.1 Concentrations of dissolved CH₄ in tropical peatlands

In contrast to the CH₄ fluxes, dissolved CH₄ concentrations were mainly affected by land use (**Fig. 6.5**). Ueda *et al.* (2000) found that dissolved CH₄ concentrations in groundwater in a coastal peat swamp in Thailand varied from 0.01 to 417 μmol L⁻¹, with average values of 48 and 226 μmol L⁻¹ in wet seasons spanning 4 years at two sampling sites. Koschorreck (2000) reported that CH₄ in pore water at the top 8 cm of a silty loam sediment on an island in the Amazon River ranged from 0 to 900 μmol L⁻¹. Terazawa *et al.* (2007) found that dissolved CH₄ concentrations in groundwater ranged from 5.6 to 28.4 μmol L⁻¹ in a floodplain forest located in northern Japan. Pangala *et al.* (2013) recently reported that dissolved CH₄ concentrations observed in the vicinity of the flooded forest sites in this study ranged from 113 to 1539 μmol L⁻¹, much higher than our results. Several differences between the two studies, including sampling depths (50 to 150 cm in Pangala *et al.*, 2013 compared to 20 to 80 cm in this study) and procedures may have resulted in this discrepancy in dissolved CH₄ concentrations. However, the primary reason for this difference is unclear.

In this study, dissolved CH₄ concentrations were highest in the flooded burnt sites (124±84 μmol L⁻¹) followed by the drained burnt site (45.2±29.8 μmol L⁻¹), the flooded forest sites (1.15±1.38 μmol L⁻¹), and the drained forest site (0.860±0.819 μmol L⁻¹). One of the possible reasons of such high dissolved CH₄ in burnt sites is the lack of large trees which have thick and deep root systems. This situation inhibited the entry of O₂ to the soil profiles. Previous studies have also reported increased dissolved CH₄ as well as decreased DO (Liebner *et al.*, 2012), redox potential (Fritz *et al.*, 2011; Koschorreck, 2000), and root density (Fritz *et al.*, 2011) in saturated peat profiles.

DO concentrations were influenced by land use only (**Table 4.3**) and were significantly higher in the forest sites (6.9±5.6 μmol L⁻¹) compared to the burnt sites (4.0±2.9 μmol L⁻¹) (**Table 4.2**). This suggested that CH₄ produced in the peat might be oxidized by plant-mediated oxygen supply in the forest sites. O₂ supply via plant roots would reduce net CH₄ production in the forest sites even under flooded conditions. The decline in DO by CH₄ oxidation could be compensated by this plant-mediated O₂ supply. The increase in the concentration of dissolved CH₄ concentrations with depth in the flooded forest sites suggested that CH₄ oxidation rate by plant-derived O₂ may depend on the amount of plant root biomass, which usually decreases with increasing depth.

The CH₄ flux in the drained forest (FD) site was the lowest among the sites (**Table 5.1 and Fig. 6.5**), likely because the FD site had the lowest water table level (–20 to –60 cm below the ground surface; **Fig. 4.5**). Under such aerobic conditions, methanotrophic activity would have been promoted over methanogenesis. Hanson and Hanson (1996) indicated that anoxic soils produce CH₄, while well-drained soils act as a sink for atmospheric CH₄ due to CH₄ oxidation. In the drained burnt site, CH₄ fluxes were lower than in the flooded forest sites, although dissolved CH₄ concentrations were much higher in the burnt sites compared to the forest sites (**Fig. 6.5**). Previous studies suggested that CH₄ diffusing toward the atmosphere is oxidized to CO₂ by methanotrophic bacteria when oxic conditions are present in the upper peat profile (Couwenberg *et al.*, 2010; Inubushi *et al.*, 2003; Jauhiainen *et al.*, 2005 and 2008). In the drained burnt site, the occasionally non-flooded condition (**Fig. 4.5**) would create an aerobic layer near the ground surface. This would allow CH₄ produced in deeper peat layers to be oxidized and CH₄ emissions and dissolved CH₄ concentrations at a depth of 20 cm to decrease. However, in the flooded burnt sites, the lack of an aerobic surface layer prohibited CH₄ oxidation, resulting in a higher CH₄ flux compared to other sites and a higher dissolved CH₄ concentration at 20 cm compared to deeper layers. Nouchi and Mariko (1993) reported that CH₄ emission rate was proportional to pore water CH₄ concentration, so the large CH₄ emissions and high dissolved CH₄ concentrations were mutually consistent.

6.4.2 Concentrations of dissolved CO₂ in tropical peatlands

The dissolved CO₂ concentrations were significantly higher in the drained forest site at depths of 40, 60, and 80 cm. These results were correspondent with the enhanced CO₂ emission by aerobic peat decomposition and plant respiration in the drained forest site (**Table 5.2**) under low water table levels. In addition, the water table level at this site was consistently at 20-60 cm below the ground surface, much lower than other sites (**Fig. 4.5**). In line with this study Laine and Päivänen (1992) have explained, when peatlands are drained, aerobic activity is possible even in deeper layers and the increased decomposition of the peat. Dissolved CO₂ concentrations at 20 cm were lower than those at deeper layers (**Table 6.2**), which could be attributed to diffusive CO₂ loss into the atmosphere through the unsaturated top layer in the drained forest site. According to the previous result in tropical peatlands, Ueda *et al.* (2000) reported the CO₂ concentrations in a coastal swamp system in Thailand varied widely in swamp water, river water, and groundwater: 1 to 3290 μmol CO₂ L⁻¹, similar to the range in this study. The dissolved CO₂ concentrations in the drained burnt site,

which was significantly lower than those in the drained forest site, suggested a low root respiration rate from poor vegetation cover. Insufficient drainage relative to the drained forest site may have also inhibited CO₂ production through peat decomposition would suppress CO₂ emissions and result in lower dissolved CO₂ concentrations. In the flooded burnt sites, a continuously flooded condition (**Fig. 4.5**) would inhibit CO₂ diffusion into the atmosphere and cause the high dissolved CO₂ concentrations at 20 cm.

6.4.3 Concentrations of dissolved N₂O in tropical peatlands

According to Ueda *et al.* (2000) the dissolved N₂O concentrations in the groundwater of a coastal peat swamp in Thailand varied from 0 to 0.012 μmol L⁻¹, with average values of 0.006 and 0.008 μmol L⁻¹ (in the wet seasons of 4 consecutive years) at two sampling sites. These results were similar to those observed in this study.

Saturated condition and less vegetation cover inhibited the entry of O₂ to soil profile, which is favorable for denitrification process. In the flooded sites, the denitrification is a major N₂O production process. Clear relationships between increasing concentrations of N₂O and decreasing availability of O₂ in the soil air have been often shown (Burford *et al.*, 1981; Dowdell and Smith, 1974; Eggington and Smith, 1986; Hansen *et al.*, 1993; Parkin and Meisinger, 1989; Rolston *et al.*, 1976). Some studies found no or only weak positive correlations of the N₂O concentration in the soil air and the N₂O emissions (Ball *et al.*, 1997; Clayton *et al.*, 1994; Goodroad and Keeney, 1985; Lessard *et al.*, 1996; Velthof *et al.*, 1996).

In this study, however, dissolved N₂O concentration at the flooded sites was lower than the drained forest site. Flooding of soils usually enforces denitrification, but N₂O flux was negative (sink) in the flooded sites. This could be probably due to denitrification which is involving the reduction of N₂O to N₂. Under saturated conditions, a temporary use of atmospheric N₂O as an electron acceptor by microorganism probably occurred under anoxic conditions. It was explained that at the higher water table level, almost no N₂O emitted since most of N₂O had been converted microbiologically into N₂ during the transport (Kliewer and Gilliam, 1995).

Nitrification is also an important microbial N process that produces N₂O. In this study, dissolved N₂O was significantly higher in drained forest sites than other sites at depths of 20 and 40 cm. In addition, N₂O flux at this site was positive (source; **Table 5.3**). It is caused by the lower water table level (aerobic) consistently kept at -20 to -60 cm (**Fig. 4.5**) and high DO concentrations (**Table 4.3**), which created slightly oxic conditions in the peat profile, favorable for N₂O production by nitrification. Bollmann and Conrad (1998) suggested that

the main source of N₂O was through nitrification when soil moisture is low and through denitrification when soil moisture is high. Burgin and Groffman (2012) found that N₂O production in intact soil cores collected from a riparian wetland in the northeastern US increased with increasing O₂ concentration. Therefore, the high concentrations of dissolved N₂O in the drained forest site may have been due to nitrification in the drained surface layer.

6.5 Summary

The investigations of field study showed that: 1) dissolved CH₄ concentrations were significantly affected by land use and drainage, which were highest in the flooded burnt sites (124±84 μmol L⁻¹) followed by the drained burnt site (45.2±29.8 μmol L⁻¹), the flooded forest sites (1.15±1.38 μmol L⁻¹), and the drained forest site (0.860±0.819 μmol L⁻¹), 2) dissolved CO₂ concentrations were 0.6 to 3.5 mmol L⁻¹, which was highest in the drained forest site, and 3) dissolved N₂O concentrations were 0.005 to 0.22 μmol L⁻¹ but occurred at <0.01 μmol L⁻¹ in most cases.

GENERAL DISCUSSION

7.1 Natural peat swamp forest and their contributions on C balance

Tropical peat swamp forest can form one of the most efficient C sequestration and storing ecosystems because it combines substantial biomass production capacity and dead biomass conservation in nutrient-poor, waterlogged soil. In this forest ecosystem, much of the carbon fixed in photosynthesis is quickly released back into the atmosphere during plant respiration, but the rest is stored in ecosystem as labile living and dead organic matter, especially as peat. Peat (and carbon) accumulates as a result of a positive net imbalance between high tropical ecosystem primary production and incomplete organic matter decomposition in permanently saturated soil conditions (Hooijer *et al.*, 2010; Wösten *et al.*, 2008).

In the flooded forest sites, the inhibition of aerobic peat decomposition would suppress CO₂ emissions (**Table 5.2**) and result in lower dissolved CO₂ concentrations (**Table 6.2; Figs. 6.6 and 6.7**) compared to those in the drained forest site. Previous studies have estimated root respirations to contribute 30 to 90% of soil CO₂ efflux (e.g., Bowden *et al.*, 1993; Epron *et al.*, 1999; Ewel *et al.*, 1987).

As with all forest, forested wetlands (swamps) are long-term sinks of atmospheric CO₂ in the absence of significant disturbances (Bridgman *et al.*, 2006; Clymo *et al.*, 1998). Decomposition of the organic matter in peat soils depends mainly on peat type, environmental condition, the decomposers present and nutrient availability (Laiho, 2006). All these factors interact, making it complicated to predict the decomposition rate. The emission of GHGs is strongly related to the decomposition rate. This study has shown that in the flooded forest sites, CO₂ fluxes were lower than in the drained forest site (**Table 5.2**). This means peat decomposition was restricted due to higher water table level in forested wetland. Soil C export from forested wetlands is dominated by soil respiration as CO₂ (Happell and Chanton, 1993; Krauss and Whitbeck, 2012; Yu *et al.*, 2008). The component for C export is vegetation and organism involved in organic matter biological decomposition (heterotrophic respiration, involving the loss of C as CO₂ and CH₄ by organism involved in aerobic and anaerobic decomposition of organic matter, comprising plant litter, roots, and their exudates, dead animals, fungi, bacteria and the peat. While, smaller amount of C is leached out from the system in drainage run off as dissolved organic C (DOC) or particulate organic C (POC; Moore *et al.*, 2011). Previous studies have reported the autotrophic respiration usually

contribute between 35 and 50% to ecosystem respiration (Crow and Wieder 2005; Silvola *et al.*, 1996). The higher CO₂ fluxes in the drained forest site probably due to the increase in peat decomposition under lower water table level condition.

In a natural state, peat swamp forest are characterized by dense forest vegetation, with a thickness varying from 0.3 to 20 m and a groundwater table that is at or close the peat surface throughout the year (Anderson, 1983; Hirano *et al.*, 2009; Page *et al.*, 2004; Takahashi *et al.*, 2002; Wösten *et al.*, 2008). The peat accumulation rates can be rapid, averaging 4–5 mm yr⁻¹, with some sites as fast as 5–10 mm yr⁻¹ (Maas, 1996). These rates are significantly faster than in most temperate and boreal peatlands, which accumulate less than 1 mm yr⁻¹, often slower than 0.5 mm yr⁻¹ (Gorham, 1991; Gorham *et al.*, 2003).

Trees in wetlands have the capacity to cope with soil anoxia through development of morphological adaptations such as hypertrophied lenticels, adventitious roots and enlarged aerenchyma. These structures promote gas exchange between the atmosphere and the rhizosphere (Megonigal and Day, 1992; Kozłowski, 1997), in particular, entry of O₂ to the root zone. According to a recent review by Laanbroek (2010), the percentage of CH₄ oxidized before entering the atmosphere ranges from 0 to up to >90% of the potential CH₄ efflux. Up to date, the majority of tree species that possess adaptive structures to facilitate O₂ ingress are also capable of mediating CH₄ egress. According to previous studies, six tree species in Borneo Island have been confirmed to emit CH₄ from their stems: *Elaeocarpaceae* (*Elaeocarpus mastersii*), *Ebenaceae* (*Diospyros bantamensis*), *Myrtaceae* (*Tristaniopsis* sp. 2), *Clusiaceae* or *Guttiferae* (*Mesua* sp. 1), *Lauraceae* (*Litsea elliptica*), and *Annonaceae* (*Xylopiacusca*) (Pangala *et al.*, 2013).

The low CH₄ fluxes in flooded forest sites (**Tables 5.1 and 5.4**) due to the dense vegetation in the forest sites contributed on the concentrations of DO (**Table 4.2**). Vascular plants can enhance CH₄ emission rates by serving as a conduit for gas by means of aerenchyma system by which CH₄ produced in anaerobic environments bypasses the oxidation zone, where less than 10% of CH₄ may diffuse across the oxic line (unsaturated zone) (Frenzel *et al.*, 1990). Thereby, plant-mediated transport is known to be of the most importance, accounting for up to 90% of total CH₄ emission (Banker *et al.*, 1995; Holzapfel-Pschorn *et al.*, 1986; Jia *et al.*, 2001; Yagi *et al.*, 1996). Besides transporting CH₄, plants can also stimulate CH₄ production by providing substrates for methanogens (Lai, 2009; Whalen, 2005) and facilitate CH₄ oxidation by delivering O₂ downward into the rhizosphere for methanotrophs (Jia *et al.*, 2001; van der Nat and Midderberg, 1998), and also transport labile

carbon compound into anoxic soil layers (Chanton *et al.*, 2008; Joabsson and Christensen, 2001; Ström *et al.*, 2003).

The low N₂O fluxes and dissolved N₂O concentrations in these study sites were probably due to the N₂O formation in soils, which is predominantly via denitrification in anaerobic micro sites, brought about by the inhibition of aeration at high water contents (Davidson and Verchot, 2000; Linn and Doran, 1984; Smith *et al.*, 1998). At higher water table level, denitrification processes becomes dominated.

7.2 GHG fluxes from peat swamp forest affected by artificial drainage

Since the 1970s large areas of lowland tropical peatland in Southeast Asia have been converted to agriculture. Development of peatland usually accompanied the drainage of water from that areas, so much of the carbon has been lost by aerobic peat decomposition. In addition, the water level was likely to become uncontrollable during an exceptionally long dry period, which was frequently observed in El Niño and Southern Oscillation (ENSO) years (Hirano *et al.*, 2007; 2012; van der Werf *et al.*, 2008), and to make the peat severely dry and susceptible to fire. In fact, decrease in water table level enhances the oxidation rate of peat, due to increases in O₂ diffusion into the subsurface layer through the air-filled pore space. The higher CO₂ fluxes in the drained forest site (**Table 5.2**) indicated the canal constructions could affect the GHG dynamics by decreasing water table level in the tropical peatland areas.

The high CO₂ fluxes in the drained forest site suggested that CO₂ production was enhanced due to root respiration at this site. In addition, the water table level at this site was consistently at 20 to 60 cm below the ground surface, much lower than the other sites (**Fig. 4.5**). The aerobic conditions that resulted from the low water table level in the drained forest site should promote peat decomposition and contribute to high CO₂ emissions at this site. Increased CO₂ emissions at lower water table levels are attributed to increased O₂ diffusion into unsaturated peat (Clymo, 1983), aerobic respiration being more efficient in CO₂ production than anerobic respiration (Schlesinger, 1997), and increased transportation of CO₂ through unsaturated peat (Moore and Dalva, 1993).

In laboratory incubations, low water table treatments increased CO₂ emissions 2–4 times higher than those from saturated soil treatments (Freeman *et al.*, 1993c; Funk *et al.*, 1994; Moore and Dalva, 1993; Moore and Knowles, 1989). Field studies have also measured higher CO₂ emissions at lower water table levels (Chimner, 2000; Kim and Verma, 1992; Silvola *et al.*, 1996). Increased CO₂ emissions at lower water table levels are attributed to increased O₂ diffusion into unsaturated peat (Clymo, 1983). When peatlands are drained,

aerobic activity becomes possible even in deeper layers and the increased decomposition of the peat will increase CO₂ fluxes to the atmosphere (Laine and Päivänen, 1992). Drainage to greater depths increases the area of profile oxidation and, thus, tends to increase C loss (Armentano and Menges, 1986; Furukawa *et al.*, 2005).

In addition, peatland drainage for cultivation can even be net sinks of CH₄ due to oxidation of CH₄ to CO₂ in the upper aerated peat layers (Maljanen *et al.*, 2003a). The availability of O₂ is important for heterotrophic and autotrophic respiration, as the rate in oxic zones is considerably higher than in anoxic zones (Öquist and Sundh, 1998; Updegraff *et al.*, 1995). The low CH₄ fluxes in the drained sites (**Table 5.1**) can be attributed to their oxic conditions. Drainage and subsequent oxidation of wetland profiles can decrease CH₄ production (Wang *et al.*, 1996). In addition, soil-atmosphere CH₄ exchange is the result of simultaneously occurring production and consumption processes in soils, and is thus controlled by CH₄-producing methanogens operating at aerobic conditions and CH₄-consuming methanotrophs that depend on O₂ as a terminal electron acceptor (Topp and Pattey, 1997). Water table exerts a major control on CH₄ flux from wetlands through its effect on the creation of aerobic and anaerobic zones in the soil profile, and thus the potential for CH₄ consumption and production (Moore and Roulet, 1995).

In the drained sites, peatland might be a source of atmospheric N₂O due to nitrification processes (**Table 5.3; Fig. 5.6**). The averaged N₂O flux was positive in the drained sites and negative in the flooded sites, although the differences were not significant. Comparing the negative N₂O fluxes between different land uses, the flooded burnt sites consumed more N₂O than the flooded forest sites. This could be attributed to the DO concentration in the forest sites were significantly higher than in the burnt sites.

7.3 GHG fluxes from degraded peatland affected by deforestation and fire

Large areas of tropical peatlands have been damaged by forest fires and deforestation in Indonesia (Muhammad and Rieley, 2002; Page *et al.*, 2002). In 1995, more than 1 Mha of tropical peatland in Central Kalimantan in Indonesia was reclaimed for agricultural development by the “Mega Rice Project” (Muhammad and Rieley, 2002). Deforestation and fire could be changed physical and chemical properties in soils that influence on dynamics of GHGs. In Southeast Asia, repeated fire events are common on drainage-affected peatland. Each fire event has caused successive losses of vegetation (Hoscilo *et al.*, 2011). As a result of peat combustion, older and more recalcitrant peat was exposed (Page *et al.*, 2002). Degradation of tropical peatlands leads to release of C and a reduction in the size of their C

stores (Hooijer *et al.*, 2006 and 2010; Jauhiainen *et al.*, 2005 and 2008; Page *et al.*, 2002; Rieley *et al.*, 2008). The large-scale degradation and relatively large amount of CO₂ have been caused by drainage and associated peat fires (Hooijer *et al.*, 2006; Page *et al.*, 2002).

Several studies have indicated that land cover change determines the overall GHG balance of the landscape (Borken and Beese, 2005 and 2006; Werner *et al.*, 2006; Bréchet *et al.*, 2009). Less vegetation covers and saturated conditions in wet season inhibit O₂ diffusion in tropical peatland. Plant species greatly affect the magnitude of CH₄ emissions from wetlands (Crillet *et al.*, 1988; Whalen and Reeburgh, 1988), especially when plant species varied from bryophytes, which lack vascular structure, to vascular plants (Bubier, 1995). Trees also have the capacity to cope with soil anoxia through development of morphological adaptations such as hypertrophied lenticels, adventitious roots, and enlarged aerenchyma. These structures promoted gas exchange between the atmosphere and the rhizosphere (Kozłowski, 1997; Megonigal and Day, 1992), in particular entry of O₂ to the root zone. Recent studies have demonstrated that temperate zone and tropical trees adapted to wet soil also facilitate egress of soil-produced CH₄ (Gauci *et al.*, 2010; Rise *et al.*, 2010; Pangala *et al.*, 2013; Rusch and Rennenberg, 1998; Terazawa *et al.*, 2007; Vann and Megonigal, 2003). Thus, Verville *et al.* (1998) has described on a local scale, vegetation composition had a greater effect on CH₄ emissions than direct manipulation of air and soil temperature. As shown by **Table 5.1** the CH₄ fluxes in the natural forest sites were significant lower compared than the burnt sites.

In the flooded burnt sites, a continuously flooded condition (**Fig. 4.5**) would be enhanced CH₄ production, and contributed in higher CH₄ fluxes to the atmosphere (**Table 5.1; Fig 5.2**). It could be suggested that the low density of vegetation after fire enhanced the production of CH₄ in the soils. When there is no plant, the soil would produce more CH₄ due to low DO concentrations (anaerobic conditions). CH₄ is known to be the end product of organic matter biodegradation in various anaerobic environments (Cicerone *et al.*, 1983). In addition, CH₄ production potential exponentially increased with lowering of redox potential (Yu *et al.*, 2001).

The CO₂ fluxes in the drained burnt site, which was significantly lower than those in the drained forest site (**Table 5.4; Fig. 5.4**), suggested a low root respiration rate from poor vegetation cover. Then the insufficient drainage relative to the drained forest site (**Fig. 4.5**) may have also inhibited CO₂ production through peat decomposition in the drained burnt site. Root and rhizosphere respiration is another important component of soil CO₂ flux. Pietikäinen *et al.* (1999) and Widén and Majdi (2001) found the highest respiratory activities

in boreal forest in organic layer close to the soil surface whereby the amount of fine root biomass was also the highest.

In this study, NO_3^- concentrations in peat water were slightly higher in the burnt sites than in the forest sites (**Table 4.5**). However, the effect of this NO_3^- differences on N_2O flux was unclear, probably because the NO_3^- concentration was too small to influence N_2O emissions.

7.4 Effect of land use change and drainage conditions on C balance in the peatlands

Peatland has the potential to mediate strong feedbacks on the global climate system, so their response to future climate change is important. Extensive peatland areas in Southeast Asia have been degraded through deforestation, drainage, and fire, leading to on- and off- site environmental and socio-economic impacts of local to global significance. Re-wetting the peat is an important key to vegetation restoration and protection of remaining peat carbon stock. The effectiveness of hydrological restoration is discussed and likely impacts on GHG emissions evaluated. Initial results indicate that raised water level have limited short-term impact on reduction CO_2 emissions, but could be critical in reducing fire risk (Page *et al.*, 2009). Hydrological restoration of peat may not instantly effect the reduction of peat GHG fluxes due to difficulties in creating near peat surface water table level. If only water table level is recovered without vegetation recovery, the potential CH_4 fluxes will increase due to anaerobic processes in peat under saturated conditions, especially in flooded burnt sites. Based on the results of this study, the reclamation of peatland has resulting in an abrupt and permanent shift in the ecosystem C balance from sink to source. In drained conditions CO_2 fluxes will increase due to lower table level and enhanced peat decomposition. Also in the drained sites, peat becomes a source of N_2O and it contributed increase N_2O fluxes to the atmosphere. The changes of GHG fluxes also will increase GWP from peatland due to land use change. The natural conditions have a better ability in C cycles compared the artificial conditions. In the hydrological restoration program it should not just recovery the water table level but it is better could be accompanied with reforestation or revegetation, especially in the degraded land, which affected by deforestation or fire.

CONCLUSIONS

In conclusion, the results of this study have shown the underlying processes that produce and consume the GHGs are affected by the land use change and drainage conditions. The effect of land use change and drainage conditions were significantly influences the magnitude of soil GHGs. The differences in the tropical peatlands environment (flooded and drained conditions) at this study sites, may lead to differences in C cycling and GHG production. In the relationships of DO and GHGs dynamics, especially CH₄ dynamics at the vegetated and non-vegetated lands in tropical peatland, the following hypotheses are addressed: 1) DO supplied from plant roots should oxidize CH₄ produced in saturated tropical peat soil, and 2) this CH₄ oxidation should reduce total GWP in a tropical peat swamp forest. The first hypothesis was supported given that 1) CH₄ emissions in the flooded burnt sites were significantly larger than those in the flooded forest sites, 2) dissolved CH₄ concentrations in the burnt sites were much higher than those in the forest sites, and 3) DO concentrations in the forest sites were significantly higher than those in the burnt sites. The CH₄ flux and the dissolved CH₄ concentrations at a depth of 20 cm in the drained burnt site were similar to values observed in the forest sites, suggesting that CH₄ oxidation in the surface soil layer occurred. The second hypothesis was weakly supported by the observed GWPs in the flooded burnt sites, which were 20% higher than those in the flooded forest sites. In this study, however, GWP was mainly determined by CO₂ flux. Consequently, GWP and CO₂ flux in the drained forest site were the highest for all study sites. N₂O flux made little contribution to GWP.

In the relationship with the hypotheses of this study, because methanogenesis is anaerobic process, peatland drainage would be expected to decrease CH₄ production rates (sink). But on the other hand, drained peatland enhances aerobic decomposition process, which increased the CO₂ fluxes to the atmosphere (source). Thus, soil also becomes as a source for N₂O fluxes because the increased of N mineralization. When the natural conditions of the peatland are so disrupted by drainage, oxidation and fire have caused there is no prospect of restoring after reclamation. In relations with self-sustaining living peat, the focus must be on slowing down peat oxidation and preventing fire by re-wetting the peatland as completely as possible. The challenge for such severely degraded areas is to develop and implement land use systems that combine economic benefits with maximal re-wetting, but also should be noted, because these areas have the potentially increase CH₄ fluxes to

atmosphere. That is why in the future, the management of tropical peatlands, especially in the hydrological restoration program it should not just recover the water table level but it is better could be accompanied with reforestation or revegetation, especially in the degraded land, which affected by deforestation or fire.

References

- Aber JD, Goodale CL, Ollinger SV, Smith ML, Magill AH, Martin ME, Hallett RA, Stoddard JL 2003: Is nitrogen deposition altering the nitrogen status of northeastern forest? *BioScience*, **53**, 375-389.
- Adamsen APS, King GM 1993: Methane consumption in temperate and sub arctic forest soil: rates, vertical zonation, and responses to water and nitrogen. *Applied and Environmental Microbiology*, **59**, 485-490.
- Aerts R, Ludwig F 1997: Water-table changes and nutritional status affect trace gas emissions from laboratory columns of peatland soils. *Soil Biology and Biochemistry*, **29**, 1691-1698.
- Alberto MCR, Arah JR M, Neue HU, Wassmann R, Lantin RS, Aduna JB, Bronson KF 2000: A sampling technique for the determination of dissolved methane in soil solution. *Chemosphere: Global Change Science*, **2**, 57-63.
- Albrecht B, Junge C, Zakosek H 1970: Der N₂O-gehalt der bodenluft in drei bodenprofilen. In: Heincke M, Kaupenjohann M 1999: Effects of soil solution on the dynamics of N₂O emissions: a review. *Nutrient Cycling in Agroecosystems*, **55**, 133-157.
- Amaral JA, Knowles R 1998: Inhibition of methane consumption in forest soils by monoterpenes. *Journal of Chemical Ecology*, **24**, 723-734.
- Ambus P, Christensen S 1994: Measurement of N₂O from a fertilized grassland: an analysis of spatial variability. *Journal of Geophysical Research*, **99**, 16549-16555.
- Ambus P, Christensen S 1995: Spatial and seasonal nitrous oxide and methane fluxes in Danish forest-, grassland-, and agroecosystems. *Journal of Environmental Quality*, **24**, 993-1001.
- Anderson IC, Poth M, Homstead J, Burdige D 1993: A comparison of NO and N₂O production by the autotrophic nitrifier *Nitrosomonas europaea* and heterotrophic nitrifier *Alcaligenes faecalis*. *Applied and Environmental Microbiology*, **59**, 3525-3533.
- Anderson JAR 1983: The tropical peat swamps of western Malesia. In: Gore AJP (ed) *Ecosystem of the world 413, mires, swamp, bog, fen and moor, regional studies*. pp. 181-199. Amsterdam, Elsevier.
- Anderson JM 1973: Carbon dioxide evolution from two temperate, deciduous woodland soils. *Journal of Applied Ecology*, **10**, 361-378.
- Andriess JP (ed) 1988: Nature and management of tropical peat soils. *FAO Soils Bulletin*, **59**, Italy, Rome.
- Arah JRM, Smith KA, Crichton IJ, Li HS 1991: Nitrous oxide production and denitrification in Scottish arable soils. *Journal of Soil Science*, **42**, 351-367.
- Armentano TV, Menges ES 1986: Patterns of change in the carbon balance of organic soil-wetlands of the temperate zone. *Journal of Ecology*, **74**, 755-774.
- Aselmann I, Crutzen PJ 1989: Global distribution of natural freshwater wetlands and rice paddies, their net primary productivity, seasonality and possible methane emissions. *Journal of Atmospheric Chemistry*, **8**, 307-358.
- Badan Pusat Statistik (BPS) 2012: Statistik Kependudukan Kalimantan Tengah 2012. Badan Pusat Statistik-Kalimantan Tengah. 58 hal.

- Ball BC, Horgan GW, Clayton H, Parker JP 1997: Spatial variability of nitrous oxide fluxes and controlling soil and topographic properties. *Journal of Environmental Quality*, **5**, 1399-1409.
- Banker BC, Kludze HK, Alford DP, DeLaune RD, Lindau CW 1995: Methane sources and sinks in paddy rice soils: relationship to emissions. *Agriculture Ecosystems and Environment*, **53**, 243-251.
- Bartlett KB, Crill PM, Sebacher DI, Harriss RC, Wilson JO, Melack JM 1988: Methane flux from the Central Amazonian floodplain. *Journal of Geophysical Research*, **93**, 1574-1582.
- Bartlett KB, Harriss RC 1993: Review and assessment of methane emissions from wetlands. *Chemosphere*, **26**, 261-320.
- Bastviken D, Ejlertsson J, Tranvik L 2002: Measurement of methane oxidation in lakes: a comparison of methods. *Environmental Science and Technology*, **36**, 3354-3361.
- Beauchamp EG 1997: Nitrous oxide emission from agricultural soils. *Canadian Journal of Soil Science*, **77**, 113-123.
- Belanger SE, Farris JL, Cherry DS, Cairns J Jr 1985: Sediment preference of the freshwater Asiatic clam, *Corbicula fluminea*. *Nautilus*, **99**, 66-73.
- Bellisario LM, Bubier JL, Moore TR, Chanton JP 1999: Controls on CH₄ emissions from a northern peatland. *Global Biogeochemical Cycles*, **13**, 81-91.
- Bender M, Conrad R 1994: Methane oxidation activity in various soils and freshwater sediments: occurrence, characteristics, vertical profiles, and distribution on grain size fractions. *Journal of Geophysical Research Atmosphere*, **99**, 16531-16540.
- Bender M, Conrad R 1994: Microbial oxidation of methane, ammonium and carbon monoxide, and turnover of nitrous oxide and nitric oxide in soils. *Biogeochemistry*, **27**, 97-112.
- Benstead J, Lloyd D 1994: Direct mass spectrometric measurement of gases in peat cores. *FEMS Microbiology Ecology*, **13**, 233-240.
- Binkley D, Richter D 1987: Nutrient cycles and H⁺ budgets of forest ecosystems. *Advances in ecological research*. Vol. 16. pp. 1-51. Academic press inc, London.
- Blodau C, Moore TR 2003: Micro-scale CO₂ and CH₄ dynamics in a peat soil during a water fluctuation and sulfate pulse. *Soil Biology and Biochemistry*, **35**, 535-547.
- Blodau C, Roulet NT, Heitmann T, Stewart H, Beer J, Lafleur P, Moore TR 2007: Belowground carbon turnover in a temperate ombrotrophic bog. *Global Biogeochemical Cycles*, **21**, GB1021.
- Boeckx P, Van Cleemput O, Villaralvo I 1997: Methane oxidation in soils with different textures and land use. *Nutrient Cycling in Agroecosystems*, **49**, 91-95.
- Bollmann A, Conrad R 1998: Influence of O₂ availability on NO and N₂O release by nitrification and denitrification in soils. *Global Change Biology*, **4**, 387-396.
- Boone RD, Nadelhoffer KJ, Canary JD, Kaye JP 1998: Roots exert a strong influence on the temperature sensitivity of soil respiration. *Nature*, **396**, 570-572.
- Borken W, Beese F 2005: Soil respiration in pure and mixed stands of European beech and Norway spruce following removal of organic horizons. *Canadian Journal of Forest Research*, **35**, 2756-2764.

- Borken W, Brumme R 1997: Liming practice in temperate forest ecosystems and the effects on CO₂, N₂O, and CH₄ fluxes. *Soil Use Management*, **13**, 251-257.
- Bouwman AF 1990: Exchange of greenhouse gases between terrestrial ecosystems and the atmosphere. In: Bouman AF (ed) *Soils and the green house effect*. pp. 61-127, JohnWiley & Sons, Chichester, UK.
- Bowden RD, Castro MS, Melillo JM, Steudler PA, Aber JD 1993: Fluxes of greenhouse gases between soils and the atmosphere in a temperate forest following a simulated hurricane blowdown. *Biogeochemistry*, **21**, 61-71.
- Bowden RD, Melillo JM, Steudler PA, Aber JD 1991: Effects of nitrogen additions on annual nitrous oxide fluxes from temperate forest soils in the northeastern United States. *Journal of Geophysical Research*, **96**, 9321-9328.
- Bowden WB, Bormann FH 1986: Transport and loss of nitrous oxide in soil water after forest clear-cutting. *Science*, **233**, 867-869.
- Bréchet L, Ponton S, Roy J, Freycon V, Coûteaux M, Bonal D, Epron D 2009: Do tree species characteristics influence soil respiration in tropical forest? A test based on 16 tree species planted in monospecific plots. *Plant Soil*, **319**, 235-246.
- Bremner JM 1997: Sources of nitrous oxide in soils. *Nutrient Cycling in Agroecosystems*, **49**, 7-16.
- Bremner JM, Blackmer AM, Waring SA 1980: Formation of nitrous oxide and dinitrogen by chemical decomposition of hydroxylamine in soils. *Soil Biology and Biochemistry*, **12**, 263-269.
- Breuer L, Papen H, Butterbach-Bahl K 2000: N₂O emission from tropical forest soils of Australia. *Journal of Geophysical Research*, **105**, 26,353-26,367.
- Bridgham SD, Johnston CA, Pastor P, Updegraff K 1995: Potential feedbacks of northern wetlands on climate change. *BioScience*, **45**, 262-274.
- Bridgham SD, Megonigal JP, Keller JK, Bliss NB, Trettin C 2006: The carbon balance of north American wetlands. *Wetlands*, **26**, 889-916.
- Bridgham SD, Ricardson CJ 1992: Mechanisms controlling soil respiration (CO₂ and CH₄) in southern peatlands. *Soil Biology and Biochemistry*, **24**, 1089-1099.
- Brix H, Sorrell BK, Orr PT 1992: Internal pressurization and convective gas flow in some emergent freshwater macrophytes. *American Society of Limnology and Oceanography*, **37**, 1420-1433.
- Brooks RP, Hughes RM 1988: Guidelines for assessing the biotic communities of freshwater wetlands. pp. 276–282. In: Kusler JA, Quammen ML, Brooks G (eds) *Proceedings of the national wetlands symposium: mitigation of impacts and losses*, New Orleans, LA, October 8–10, 1986. 460 pp. Association of state wetland managers. Berne, NY.
- Brown CD, Fryer CJ, Walker A 2001: Influence of topsoil tilth and soil moisture status on losses of pesticide to drains from a heavy clay soil. *Pest Management Science*, **57**, 1127-1134.
- Brumme R, Beese F 1992: Effects of liming and nitrogen fertilization on emissions of CO₂ and N₂O from a temperate forest. *Journal of Geophysical Research*, **97**, 12851-12858.

- Brumme R, Borken W 1999: Site variation in methane oxidation as affected by atmospheric deposition and type of temperate forest ecosystem. *Global Biogeochemical Cycles*, **13**, 493-501.
- Bubier JL 1995: The relationship of vegetation to methane emission and hydrochemical gradients in northern peatlands. *Journal of Ecology*, **83**, 403-420.
- Bubier JL, Moore TR 1993: Methane emissions from wetlands in the mid boreal region of northern Ontario, Canada. *Ecology*, **74**, 2240-2254.
- Buchmann N 2000: Biotic and abiotic factors controlling soil respiration rates in *Picea abies* stands. *Soils Biology and Biochemistry*, **32**, 1625-1635.
- Burford JR, Dowdell RJ, Crees R 1981: Emission of nitrous oxide to the atmosphere from direct-drilled and ploughed clay soils. *Journal of the Science of Food and Agriculture*, **32**, 219-223.
- Burgin AJ, Groffman PM, Lewis DN 2010: Factors regulating denitrification in a riparian wetland. *Soil Science Society of America Journal*, **74**, 1826-1833.
- Burgin AM, Groffman PM 2012: Soil O₂ controls denitrification rates and N₂O yield in a riparian wetland. *Journal of Geophysical Research*, **117**, G01010.
- Butterbach-Bahl K, Breuer L, Gasche R, Willibald G, Papen H 2002: Exchange of trace gases between soils and the atmosphere in Scots pine forest ecosystems of the northeastern German lowlands: 1. Fluxes of N₂O, NO/NO₂ and CH₄ at forest sites with different N-deposition. *Forest Ecology and Management*, **167**, 123-134.
- Butterbach-Bahl K, Gasche R, Breuer L, Papen H 1997: Fluxes of NO and N₂O from temperate forest soils: impact of forest type, N deposition and of liming on the NO and N₂O. *Nutrient Cycling in Agroecosystems*, **48**, 79-90.
- Butterbach-Bahl K, Gasche R, Huber CH, Kreuzer K, Papen H 1998: Impact of N-input by wet deposition on N-trace gas fluxes and CH₄-oxidation in spruce forest ecosystems of the temperate zone in Europe. *Atmospheric Environment*, **32**, 559-564.
- Buttler AJ, Diné H, Lévesque M, Mathur SP 1991: The relation between movement of subsurface water and gaseous methane in a basin bog with a novel instrument. *Canadian Journal of Soil Science*, **71**, 427-438.
- Buyanovsky GA, Wagner GH, Gantzer CJ 1986: Soil respiration in a winter wheat ecosystem. *Soil Science Society of America Journal*, **50**, 338-344.
- Castro MS, Steudler PA, Melillo JM, Aber JD, Bowden RD 1995: Factors controlling atmospheric methane consumption by temperate forest soils. *Global Biogeochemical Cycle*, **9**, 1-10.
- Castro MS, Steudler PA, Melillo JM, Aber JD, Millham S 1993: Exchange of N₂O and CH₄ between the atmosphere and soils in spruce-fir forest in the northeastern United States. *Biogeochemistry*, **18**, 119-135.
- Cates RL, Keeney DR 1987: Nitrous oxide production throughout the year from fertilized and manured maize fields. *Journal of Environmental Quality*, **16**, 443-447.
- Chapin III FS, Ruess RW 2001: The roots of the matter. *Nature*, **411**, 749-752.
- Chapuis-Lardy L, Wragge N, Metay A, Chottes JL 2007: Soils, a sin for N₂O? A review. *Global Change Biology*, **13**, 1-17.

- Chimner RA 2000: Carbon dynamics of Southern Rocky Mountain fens. PhD Dissertation, Colorado State University, Ft Collins, CO. In: Chimner RA, Cooper DJ 2003: Influence of water table levels on CO₂ emissions in a Colorado subalpine fen: an in situ microcosm study. *Soil Biology and Biochemistry*, **35**, 345-351.
- Chimner RA 2004: Soil respiration rates of tropical peatland in Micronesia and Hawaii. *Wetlands*, **24**, 51-56.
- Christensen TR, Ekberg A, Ström L, Mastepanov M, Panikov N, Oquist M, Svensson BH, Nykänen NH, Martikainen PJ, Oskarsson H 2003: Factors controlling large scale variations in methane emissions from wetlands. *Geophysical Research Letters*, **30**, 1414.
- Cicerone RJ, Shetter JD 1981: Sources of atmospheric methane measurements in paddies and a discussion. *Journal of Geophysical Research*, **86**, 7203-7209.
- Clayton H, Arah JRM, Smith KA 1994: Measurement of nitrous oxide emissions from fertilized grassland using closed chambers. *Journal of Geophysical Research: Atmospheres*, **99**, 16599-16607.
- Clymo RS 1983: Peat. In: Gore AJP (ed) *Ecosystems of the world*, 4A. Mires: swamp, bog, fen, and moor. General studies. pp. 159-224. Elsevier, Amsterdam.
- Clymo RS, Bryant CL 2008: Diffusion and mass flow of dissolved carbon dioxide, methane, and dissolved organic carbon in a 7-m raised peat bog. *Geochimica et Cosmochimica Acta*, **72**, 2048-2066.
- Clymo RS, Pearce DME 1995: Methane and carbon dioxide production in, transport through, and efflux from a peatland. *Philosophical Transactions of the Royal Society London A*, **350**, 249-259.
- Clymo RS, Turunen J, Tolonen K 1998: Carbon accumulation in peatland. *Oikos*, **81**, 368-388.
- Conrad R 1995: Soil microbial process involved in production and consumption of atmospheric trace gases. *Advances in Microbial Ecology*, **14**, 207-250.
- Conrad R 1996: Soil microorganisms as controllers of atmospheric trace gases (H₂, CO, CH₄, OCS, N₂O, and NO). *Microbiological reviews*, **60**, 609-640.
- Conrad R 2002: Microbiological and biochemical background of production and consumption of NO and N₂O in soils. In: Gasche R, Papen H, Rennenberg H (eds) *Trace gas exchange in forest ecosystems*. pp. 3-33. Kluwer academic publisher.
- Couwenberg J 2009: Methane emissions from peat soils. (organic soils, histosols): facts, MRV-ability, emissions factors. 14 pp. Wetlands international.
- Couwenberg J, Domain R, Joosten H 2010: Greenhouse gas fluxes from tropical peatlands in South-East Asia. *Global Change Biology*, **16**, 1715-1732.
- Crawford RMM 1992: Oxygen availability as an ecological limit to plant distribution. In: Begon M, Fitter AH (eds) *Advances in ecological research*, **23**, 93-185.
- Crill PM 1991: Seasonal patterns of methane uptake and carbon dioxide release by a temperate woodland soil. *Global Biogeochemical Cycle*, **5**, 319-334.
- Crill PM, Bartlett KB, Harriss RC, Gorham E, Verry ES, Sebacher DI, Madzar L, Sanner W 1988: Methane flux from Minnesota peatlands. *Global Biogeochemistry Cycles*, **2**, 371-384.
- Crow SE, Wieder RK 2005: Sources of CO₂ emission from a northern peatland: root respiration, exudation, and decomposition. *Ecology*, **86**, 1825-1834.

- Crutzen PJ 1970: The influence of nitrogen oxides on the atmospheric ozone content. *Quarterly Journal Royal Meteorological Society*, **96**, 320-325.
- Crutzen PJ, Ehhalt DH 1977: Effects of nitrogen fertilizers and combustion on the stratospheric ozone layer. *Ambio*, **6**, 112-117.
- Curl EA, Truelove B 1986: *The rhizosphere*. 290 pp. Springer-Verlag, Berlin, Germany.
- Daulat WE, Clymo RS 1998: Effects of soil temperature and water table on the efflux of methane from peatland surface cores. *Atmospheric Environment*, **32**, 3207-3218.
- Davidson EA 1991: Fluxes of nitrous oxide and nitric oxide from terrestrial ecosystems. In: Rogers JE, Whitman WB (eds) *Microbial production and consumption of greenhouse gases: methane, nitrogen oxides, and halomethanes*. pp. 219-235. American Society for Microbiology. Washington DC, USA.
- Davidson EA 1992: Pulses of nitric oxide and nitrous oxide flux following wetting of dry soil: an assessment of probable sources and importance relative to annual fluxes. *Ecological Bulletin*, **42**, 149-155.
- Davidson EA, Belk E, Boone RD 1998: Soil water content and temperature as independent or confounded factors controlling soil respiration in a temperate mixed hardwood forest. *Global Change Biology*, **4**, 217-227.
- Davidson EA, Firestone MK 1988: Measurement of nitrous oxide dissolved in soil solution. *Soil Science Society of America Journal*, **52**, 1201-1203.
- Davidson EA, Hart SC, Shanks CA, Firestone MK 1991: Measuring gross nitrogen mineralization, and nitrification by ¹⁵N isotopic pool dilution in intact soil cores. *Soil Science*, **42**, 335-349.
- Davidson EA, Kinglerlee W 1997: A global inventory of nitric oxide emissions from soils. *Nutrient Cycling in Agroecosystems*, **48**, 37-50.
- Davidson EA, Matson PA, Vitousek PM, Riley R, Dunkin K, Garcia-Mendez G, Maass JM 1993: Processes regulating soil emissions of NO and N₂O in a seasonally dry tropical forest. *Ecology*, **74**, 130-139.
- Davidson EA, Swank WT 1987: Factors limiting denitrification in soils from mature and disturbed southeastern Hardwood forests. *Forest Science*, **33**, 135-144.
- Davidson EA, Verchot LV 2000: Testing the hole-in the pipe model of nitric and nitrous oxide emissions from soils using the TRAGNET database. *Global Biogeochemical Cycles*, **14**, 1035-1043.
- De Boer, Gunnewiek PAK, Laanbroek HJ 1995: Ammonium-oxidation at low pH by a chemolithotrophic bacterium belonging to the genus *Nitrosospira*. *Soil Biology and Biochemistry*, **27**, 127-132.
- De la Cruz AA 1986: Tropical wetlands as a carbon source. *Aquatic Botany*, **25**, 109-115.
- Delf Hydraulics 2006: Peat-CO₂, assessment of CO₂ emissions from drained peatlands in SE Asia. Report R & D Projects Q3943/Q3684/Q4142. 1st ed. December 7, 2006.
- Delmas RA, Servant J, Tathy JP, Cros B, Labat M 1992: Sources and sinks of methane and carbon dioxide exchanges in mountain forest in equatorial Africa. *Journal of Geophysical Research*, **97**, 6169-6179.
- Denmead OT, Freney JR, Simpson JR 1979: Studies of nitrous oxide emission from a grass sward. *Soil Science Society of America Journal*, **43**, 726-728.

- Ding W, Cai Z, Tsuruta H 2004: Methane concentration and emission as affected by methane transport capacity of plants in freshwater marsh. *Water, Air, and Soil Pollution*, **158**, 99-111.
- Dise NB 1993: Methane emission from Minnesota peatlands: spatial and seasonal variability. *Global Biogeochemical Cycles*, **7**, 123-142.
- Dobbie KE, McTaggart IP, Smith KA 1999: Nitrous oxide emissions from intensive agriculture systems: variations between crops and seasons, key driving variables, and mean emission factors. *Journal of Geophysical Research*, **104**, 26891-26899.
- Dobbie KE, Smith KA 1996: Comparison of CH₄ oxidation rates in woodland, arable and set aside soils. *Soils Biology and Biochemistry*, **28**, 1357-1365.
- Dong Y, Scharffe D, Lobert JM, Crutzen PJ, Sanhueza E 1998: Fluxes of CO₂, CH₄ and N₂O from a temperate forest soil: the effects of leaves and humus layers. *Tellus*, **50B**, 243-252.
- Dorr H, Katruff L, Levin I 1993: Soil texture parameterization of the methane uptake in aerated soils. *Chemosphere*, **26**, 697-713.
- Dorran JW, Mielke LN, Power JF 1990: Microbial activity as regulated by soil water-filled pore space. In: Davidson EA, Belk E, Boone RD 1998: Soil water content and temperature as independent or confounded factors controlling soil respiration in a temperate mixed hardwood forest. *Global Change Biology*, **4**, 217-227.
- Dowdell RJ, Burford JR, Cress R 1979: Losses of nitrous oxide dissolved in drainage water from agricultural land. *Nature*, **278**, 342-343.
- Dowdell RJ, Smith KA 1974: Field studies of the soil atmosphere II. Occurrence of nitrous oxide. *European Journal of Soil Science*, **25**, 231-238.
- Drew MC 1990: Sensing soil oxygen. *Plant, Cell and Environment*, **13**, 681-693.
- Drew MC, Greg Cobb B, Johnson JR, Andrews D, Morgan PW, Jordan W, He CJ 1994: Metabolic acclimation of root tips to oxygen deficiency. *Annals of Botany*, **74**, 281-286.
- Drexler JZ, de Fontaine CS, Deverel SJ 2009: The legacy of wetland drainage on the remaining peat in the Sacramento – San Joaquin delta, California, USA. *Wetlands*, **29**, 372-386.
- Dunfield P, Knowles R 1995: Kinetics of inhibition of methane oxidation by nitrate, nitrite, and ammonium in a Humisol. *Applied and Environmental Microbiology*, **61**, 3129-3135.
- Duxbury JM, Mosier AR 1993: Status and issues concerning agricultural emissions of greenhouse gases. In: Kaiser HM, Drennen TE (eds) *Agricultural dimensions of global climate change*. pp 229–258. St. Lucie press. Delray Beach, FL.
- Edwards NT 1975: Effects of temperature and moisture on carbon dioxide evolution in a mixed deciduous forest floor. *Soil Science Society of America Journal*, **39**, 361-365.
- Edwards NT, Hariss WF 1977: Carbon cycling in a mixed deciduous forest floor. *Ecology*, **58**, 431-437.
- Egginton GM, Smith KA 1986: Losses of nitrogen by denitrification from a grassland of soil fertilized with cattle slurry and calcium nitrate. *European Journal of Soil Science*, **37**, 69-80.

- Ehhalt D, Prather M, Dentener F, Derwent R, Edward JD, Holland E, Isaksen I, Katima J, Kirchhoff V, Matson P, Midgley P, Wang M, Berntsen T, Bey I, Brasseur G, Buja L, Collins WJ, Daniel JS, DeMore WB, Derek N, Dickerson R, Etheridge D, Feichter J, Fraser P, Friedl R, Fuglestvedt J, Gauss M, Grenfell L, Arnulf G, Hariss N, Hauglustaine D, *et al* 2001: Atmospheric chemistry and greenhouse gases. In: Joos F, McFarland M (eds) *Climate change 2001: Working group I: the scientific basis*. pp. 239-288.
- Eichner MJ 1990: Nitrous oxide emissions from fertilized soils: Summary of available data. *Journal of Environmental Quality*, **19**, 272-280.
- Elberling Bo, Askaer L, Jørgensen CJ, Joensen HP, Kühl M, Glud RN, Lauritsen FR 2011: Linking soil O₂, CO₂, and CH₄ concentrations in a wetland soil: implications for CO₂ and CH₄. *Environmental Science and Technology*, **45**, 3393-3399.
- EPA (Environmental Protection Agency) 2010: Methane and nitrous oxide emissions from natural source. United States Environmental Protection Agency Office of Atmospheric Programs. Washington DC. USA.
- Epron D, Farque L, Lucot E, Badot PM 1999: Soil CO₂ efflux in a beech forest: dependence on soil temperature and soil water content. *Annual Forest Science*, **56**, 221-226.
- Erickson H, Keller M, Davidson EA 2001: Nitrogen oxide fluxes and nitrogen cycling during post agricultural succession and forest fertilization in the humid tropics. *Ecosystems*, **4**, 67-84.
- Eswaran H, Van den Berg E, Reich P 1993: Organic carbon in soils of the world. *Soil Science Society of America Journal*, **57**, 192-194.
- Ewel KC, Cropper WP Jr, Gholz HL 1987: Soil CO₂ evolution in Florida slash pine plantations. II. Importance of root respiration. *Canadian Journal of Forest Research*, **17**, 330-333.
- Fang C, Moncrieff JB, Gholz HL, Clark KL 1998: Soil CO₂ efflux and its spatial variation in a Florida slash pine plantation. *Plant and Soil*, **205**, 135-146.
- FAO/ISRIC 1990: Guidelines for soil description (3rd edition, revised). Food and Agricultural Organization. Rome.
- Faulkner SP, Patrick WH 1992: Redox processes and diagnostic wetland soil indicators in bottomland hardwood forest. *Soil Science Society of America Journal*, **56**, 856-865.
- Faulkner SP, Patrick WH. Jr, Gambrell RP 1989: Field techniques for measuring wetland soil parameters. *Soil Science Society of America Journal*, **53**, 883-890.
- Fenn ME, Poth MA, Johnson DW 1996: Evidence for nitrogen saturation in the San Bernardino Mountains in southern California. *Forest Ecology and Management*, **82**, 211-230.
- Firestone MK 1982: Biological nitrification. In: Stevenson FJ (ed) *Nitrogen in agricultural soils*. pp. 289-326. Agronomy Monograph No. 22. ASA-CSSA-SSSA, Madison. WI.
- Firestone MK, Davidson EA 1989: Microbiological basis of NO and N₂O production and consumption in soil. In: Andrea MO, Schimel DS (eds) *Exchange of trace gases between terrestrial ecosystems and the atmosphere*. pp. 7-21. John Wiley & Sons. Chichester, UK.
- Flessa H, Wild U, Klemisch M, Pfadenhauer J 1998: Nitrous oxide and methane fluxes from organic soils under agriculture. *European Journal of Soil Science*, **49**, 327-335.

- Forster P, Ramaswamy V, Artaxo P, Bernsten T, Betts R, Fahey DW, Haywood J, Lean J, Lowe DC, Myhre G, Nganga J, Prinn R, Raga G, Schulz M, Van Dorland R 2007: Changes in atmospheric constituents and in radiative forcing. In: Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller BR (eds) *Climate Change 2007: The physical science basis. Contribution of working group I to the fourth assessment report of the Intergovernmental Panel on Climate Change*. pp. 500-587. Cambridge university press. Cambridge, UK and New York, US.
- Freeman C, Fenner N, Ostle NJ, Kang H, Dowrick DJ, Reynolds B, Lock MA, Sleep D, Hughes S, Hudson J 2004: Export of dissolved organic carbon from peatlands under elevated carbon dioxide levels. *Nature*, **430**, 195-198.
- Freeman C, Lock MA, Reynolds B 1993a: Climatic-change and the release of immobilized nutrients from welsh riparian wetland soils. *Ecology Engineering*, **2**, 367-373.
- Freeman C, Lock MA, Reynolds B 1993b: Impacts of climatic change on peatland hydrochemistry; a laboratory-based experiment. *Chemistry and Ecology*, **8**, 49-59.
- Freeman C, Lock MA, Reynolds B 1993c: Fluxes of CO₂, CH₄ and N₂O from a Welsh peatland following simulation of water table draw-down: potential feed back to climate change. *Biogeochemistry*, **19**, 51-60.
- Freney JR, Denmead OT, Simpson JR 1978: Soil as a source or sink for atmospheric nitrous oxide. *Nature*, **273**, 530-532.
- Frenzel P, Thebrath B, Conrad R 1990: Oxidation of methane in the oxic surface layer of deep lakes sediment (Lake Constance). *FEMS Microbiology Letters*, **73**, 149-158.
- Fritz C, Pancotto VA, Elzenga JTM, Visser EJW, Grootjans AP, Pol A, Iturraspe R, Roelofs JGM, Smolders AJP 2011: Zero methane emission bog: extreme rhizosphere oxygenation by cushion plants in Patagonia. *New Phytologist*, **190**, 398-408.
- Fuglestvedt JS, Berntsen TK, Godal O, Skodvin T 2000: Climate implication of GWP-based reductions in greenhouse gas emissions. *Geophysical Research Letters*, **27**, 409-412.
- Funk DW, Pullman ER, Peterson KM, Crill PM, Billings WD 1994: Influence of water table on carbon dioxide, carbon monoxide, and methane fluxes from taiga bog microcosms. *Global Biogeochemical Cycles*, **8**, 271-278.
- Furukawa Y, Inubushi K, Ali M, Itang AM, Tsuruta H 2005: Effect of changing groundwater levels caused by land-use changes on greenhouse gas fluxes from tropical peat lands. *Nutrient Cycling in Agroecosystems*, **71**, 81-91.
- Galchenko VF, Lein A, Ivanov M 1989: Biological sinks of methane. In: Andrea MO, Schimel DS (eds) *Exchange of trace gases between terrestrial ecosystems and the atmosphere*. pp 59-71. John Wiley & Sons. Chichester, UK.
- Galloway JN 2000: Nitrogen mobilization in Asia. *Nutrient Cycling in Agroecosystems*, **57**, 1-12.
- Galloway JN, Aber JD, Erisman JW, Seitzinger SP, Howarth RW, Cowing EB, Cosby BJ 2003: The nitrogen cascade. *BioScience*, **53**, 341-356.
- Garcia-Méndez G, Maass JM, Matson PA, Vitousek PM 1991: Nitrogen transformations and nitrous-oxide flux in a tropical deciduous forest in México. *Oecologia*, **88**, 362-366.
- Garcia-Montiel DC, Steudler PA, Piccolo MC, Melillo JM, Neill C, Cerri CC 2001: Controls on soil nitrogen oxide emissions from forest and pasture in the Brazillian Amazon. *Global Biogeochemical Cycles*, **15**, 1021-1030.

- Garrret HE, Cox GS 1973: Carbon dioxide evolution from the floor of an oak-hickory forest. *Soil Science Society of America Proceedings*, **37**, 641-644.
- Gasche R, Papen H 1999: A 3-year continuous record of nitrogen trace gas fluxes from untreated and limed soils of a N-saturated spruce and beech forest ecosystem in Germany 2. NO and NO₂ fluxes. *Journal of Geophysical Research*, **104**, 18505-18520.
- Gauci V, Gowing DJ, Hornibrook ER, Davis JM, Dise NB 2010: Woody stem methane emission in mature wetland alder trees. *Atmospheric Environment*, **44**, 2157-2160.
- Gaudinski JB, Trumbore SE, Davidson EA, Shuhui Z 2000: Soil carbon cycling in a temperate forest: radiocarbon-based estimates of residence times, sequestration rates and partitioning of fluxes. *Biogeochemistry*, **51**, 33-69.
- Germer J, Sauerborn J 2008: Estimation of the impact of oil palm plantation establishment on greenhouse gas balance. *Environment, Development and Sustainability*, **10**, 697-716.
- Gilbert B, Frenzel P 1998: Rice roots and CH₄ oxidation: the activity of bacteria, their distribution and the microenvironment. *Soil Biology and Biochemistry*, **30**, 1903-1916.
- Gödde M, Conrad R 1999: Immediate and adaptational temperature effects on nitric oxide production and nitrous oxide release from nitrification and denitrification in two soils. *Biol. Fertil. Soils*, **30**, 33-40.
- Goodroad LL, Keeney DR 1984: Nitrous oxide emission from forest, marsh, and prairie ecosystems. *Soil Science Society of America Journal*, **13**, 448-452.
- Goodroad LL, Keeney DR 1985: Site of nitrous oxide production in field soils. *Biology and Fertility of Soils*, **1**, 3-7.
- Gorham E 1991: Northern peatland: role in the carbon cycle and probable responses to climatic warming. *Ecological Applications*, **1**, 182-195.
- Gorham E, Janssens JA, Glaser PH 2003: Rates of peat accumulation during the postglacial period in 32 sites from Alaska to Newfoundland, with special emphasis on northern Minnesota. *Canadian Journal of Botany*, **81**, 429-438.
- Granli T, Boeckman OC 1994: Nitrous oxide from agriculture. *Norwegian Journal of Agricultural Sciences*, supplement **12**, 1-128.
- Groffman PM, Brumme R, Butterbach-Bahl K, Dobbie KE, Mosier AR, Ojima D, Papen H, Parton WJ, Smith KA, Wagner-Riddle C 2000: Evaluating annual nitrous oxide fluxes at the ecosystem scale. *Global Biogeochemical Cycles*, **14**, 1061-1070.
- Gulledge J, Doyle AP, Schimel JP 1997: Different NH₄⁺-inhibition patterns of soil CH₄ consumption: a result of distinct CH₄-oxidizer populations across sites? *Soil Biology & Biochemistry*, **29**, 13-21.
- Gulledge J, Schimel JP 1998: Moisture control over atmospheric CH₄ consumption and CO₂ production in diverse Alaskan soils. *Soil Biology and Biochemistry*, **30**, 1127-1132.
- Hadi A, Inubushi K, Furukawa Y, Purnomo E, Rasmadi M, Tsuruta H 2005: Greenhouse gas emissions from tropical peatlands of Kalimantan, Indonesia. *Nutrient Cycling in Agroecosystems*, **71**, 73-80.
- Hadi A, Inubushi K, Purnomo E, Razie F, Yamakawa K, Tsuruta H 2000: Effect of land-use changes on nitrous oxide (N₂O) emission from tropical peatlands. *Chemosphere Global Change Science*, **2**, 347-358.

- Hall GH, Simon BM, Pickup RW 1996: CH₄ production in blanket bog peat: a procedure for sampling, sectioning and incubating samples whilst maintaining anaerobic conditions. *Soil Biology and Biochemistry*, **28**, 9-15.
- Hall SJ, Matson PA 1999: Nitrogen oxide emissions after nitrogen additions in tropical forests. *Nature*, **400**, 152-155.
- Hampson PS 1989: Dissolved-oxygen concentrations in a central Florida wetlands stream. Proceedings of an American water resources association symposium on wetlands. Wetlands: concerns and successes, September 17-22, 1989. pp. 149-159. Tampa, Florida.
- Hansen S, Maehlum JE, Bakken LR 1993: N₂O and CH₄ fluxes in soil influenced by fertilization and tractor traffic. *Soil Biology and Biochemistry*, **5**, 621-630.
- Hanson PJ, Edwards NT, Garten CT, Andrews JA 2000: Separating root and soil microbial contributions to soil respiration: a review of methods and observations. *Biogeochemistry*, **48**, 115-146.
- Hanson PJ, Wullschleger SD, Bohlman SA, Todd DE 1993: Seasonal and topographic patterns of forest floor CO₂ efflux from an upland oak forest. *Tree Physiology*, **13**, 1-15.
- Hanson RS, Hanson TE 1996: Methanotrophic bacteria. *Microbiology and Molecular Biology Reviews*, **60**, 439-471.
- Happell JD, Chanton JP 1993: Carbon remineralization in a north Florida swamp forest: effects of water level on the pathways and rates of soil organic matter decomposition. *Global Biogeochemical Cycles*, **7**, 475-490.
- Heathwaite AL 1990: The effect of drainage on nutrient release from fen peat and its implications for water quality - a laboratory simulation. *Water, Air, and Soil Pollution*, **49**, 159-173.
- Heyer J, Berger U, Kuzin IL, Yakovlev ON 2002: Methane emissions from different ecosystem structures of the subarctic tundra in western Siberia during midsummer and during the thawing period. *Tellus*, **54B**, 231-249.
- Hirano T, Jauhiainen J, Inoue T, Takahashi H 2009: Controls on the carbon balance of tropical peatlands. *Ecosystems*, **12**, 873-887.
- Hirano T, Segah H, Harada T, Limin S, June T, Hirata R, Osaki M 2007: Carbon dioxide balance of a tropical peat swamp forest in Kalimantan, Indonesia. *Global Change Biology*, **13**, 412-425.
- Hirano T, Segah H, Kusin K, Limin S, Takahashi H, Osaki M 2012: Effects of disturbances on the carbon balance of tropical peat swamp forests. *Global Change Biology*, **18**, 3410-3422.
- Hirota M, Tang Y, Hu Q, Hirata S, Kato T, Mo W, Cao G, Mariko S 2004: Methane emissions from different vegetation zones in a Qinghai-Tibetan Plateau wetland. *Soil Biology and Biochemistry*, **36**, 737-748.
- Högberg P, Nordgren A, Buchmann N, Taylor AFS, Ekblad A, Högberg MN, Nyberg G, Ottosson-Löfvenius M, Read DJ 2001: Large-scale forest girdling shows that current photosynthesis drives soil respiration. *Nature*, **411**, 789-792.
- Holzappel-Pschorn A, Conrad R, Seiler W 1986: Effects of vegetation on the emission of methane from submerged paddy soil. *Plant and Soil*, **92**, 223-233.

- Hooijer A, Page S, Canadell JG, Silvius M, Kwadijk J, Wösten H, Jauhiainen J 2010: Current and future CO₂ emissions from drained peatlands in Southeast Asia. *Biogeosciences*, **7**, 1505-1514.
- Hooijer A, Page SE, Jauhiainen J, Lee WA, Idris A, Anshari G 2012: Subsidence and carbon loss in drained tropical peatlands. *Biogeosciences*, **9**, 1053-1071.
- Hooijer A, Silvius, M, Wösten H, Page S 2006: Peat-CO₂, assessment of CO₂ emissions from drained peatlands in SE Asia. Delft Hydraulics Report Q3943/2006, Delft, the Netherlands.
- Hoscilo A, Page SE, Rieley JO 2011: Effect of repeated fires on land-cover change on peatland in southern Central Kalimantan, Indonesia, from 1973 to 2005. *International Journal of Wildland Fire*, **20**, 578-588.
- Houghton RA 1995: Land-use change and the carbon cycle. *Global Change Biology*, **1**, 275-287.
- Houghton RA 1999: The annual net flux of carbon to the atmosphere from changes in land use 1850-1990. *Tellus*, **51B**, 298-313.
- Houghton RA, Hackler JL, Lawrence KT 1999: The U.S. carbon budget: contributions from land-use change. *Science*, **285**, 574-578.
- Houghton RA, Hobbie JE, Melillo JM, Moore B, Peterson BJ, Shaver GR, Woodwell GM. 1983: Changes in the carbon content of terrestrial biota and soils between 1860 and 1980: a net release of CO₂ to the atmosphere. *Ecological Monographs*, **53**, 235-262.
- Houghton RA, Skole DL, Nobre CA, Hackler JL, Lawrence KT, Chomentowski WH 2000: Annual fluxes of carbon from deforestation and regrowth in the Brazilian Amazon. *Nature*, **403**, 301-304.
- Howe AJ, Rodrigues, Saco PM 2009: Surface evolution and carbon sequestration in disturbed and undisturbed wetland soils of the Hunter estuary, southeast Australia. *Estuarine, Coastal and Shelf Science*, **84**, 75-83.
- Immirzi P, Maltby E, Clymo RS 1992: The global status of peatlands and their role in carbon cycling. Report no. 11. The wetland ecosystems research group. University of Exeter, UK.
- Inubushi K, Furukawa Y, Hadi A, Purnomo E, Tsuruta H 2003: Seasonal changes of CO₂, CH₄ and N₂O fluxes in relation to land-use change in tropical peatlands located in coastal area of South Kalimantan. *Chemosphere*, **52**, 603-608.
- IPCC-Intergovernmental Panel on Climate Change 1995: Climate change 1994: radiative forcing of climate change and an evaluation of the IPCC IS92 emission scenarios. In: Houghton JT, Meira Filho LG, Bruce J, Hoesung Lee, Callander BA, Haites E, Harris N, Maskell K (eds). 339 pp. Cambridge university press. Cambridge, UK and New York, USA.
- IPCC-Intergovernmental Panel on Climate Change 1996: Climate change 1995: the science of climate change. Contribution of working group I to the second assessment report of the Intergovernmental Panel on Climate Change. In: Houghton JT, Meira Filho LG, Callander BA, Harris N, Kattenberg A, Maskell K. 572 pp. Cambridge university press. Cambridge, UK and New York, USA.

- IPCC-Intergovernmental Panel on Climate Change 2001: Climate change 2001: the scientific basis. Contribution of working group I to the third assessment of the Intergovernmental Panel on Climate Change. In: Houghton JT, Ding Y, Griggs DJ, Noguer M, van der Linden PJ, Dai X, Maskell K, Johnson CA (eds). Cambridge university press. Cambridge, UK and New York, USA.
- IPCC-Intergovernmental Panel on Climate Change 2007: Summary for Policymakers. In: Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL Jr (eds) Climate change 2007: the physical science basis. Contribution of working group I to the fourth assessment report of the Intergovernmental Panel on Climate Change. pp 1-18. Cambridge university press, UK and New York, USA.
- IPCC-International Panel on Climate Change 2007: Climate change 2007: the physical science basis. Contribution of working group I to the fourth assessment report of the Intergovernmental Panel on Climate Change In: Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL Jr (eds). Cambridge university press, Cambridge and New York, USA.
- Ismail Z 1999: Survey of fish diversity in peat swamp forest. In: Yuan, CT, Havmoller P (eds) Sustainable management of peat swamp forest in Peninsular Malaysia. pp. 173–198. Forestry Department Peninsular Malaysia, Kuala Lumpur.
- Jali D 2004: Nitrogen mineralization in the tropical peat swamps. In: Jauhiainen J, Limin S, Silvennoinen H, Vasander H 2008: Carbon dioxide and methane fluxes in drained tropical peat before and after hydrological restoration. *Ecology*, **89**, 3503-3514.
- Janssens IA, Lankreijer H, Matteucci G, Kowalski AS, Buchmann N, Epron D, Pilegaard K, Kutsch W, Longdoz B, Grünwald T, Montagnani L, Dore S, Rebmanns C, Moors EJ, Grelle A, Rannik Ü, Morgenstern K, Oltchev S, Clement R, Gudmundsson J, Minerbi M, Berbigier P, Ibrom A, Moncrieff J, Aubinet M, Bernhofer C, Jensen O, Vesala T, Granier A, Schulze ED, Lindroth A, Dolman AJ, Jarvis PG, Ceulemans R, Valentini R 2001: Productivity overshadows temperature in determining soil and ecosystem respiration across European forests. *Global Change Biology*, **7**, 269-278.
- Jauhiainen J, Heikkinen J, Martikainen PJ, Vasander H 2001: CO₂ and CH₄ fluxes in pristine and peatland converted to agriculture in Central Kalimantan, Indonesia. *International Peat Journal*, **11**, 43-49.
- Jauhiainen J, Silvennoinen H, Hämäläinen R, Kusin K, Limin S, Raison RJ, Vasander H 2012: Nitrous oxide fluxes from tropical peat with different disturbance history and management. *Biogeosciences*, **9**, 1337-1350.
- Jauhiainen J, Limin S, Silvennoinen H, Vasander H 2008: Carbon dioxide and methane fluxes in drained tropical peat before and after hydrological restoration. *Ecology*, **89**, 3503–3514.
- Jauhiainen J, Takahashi H, Heikkinen JEP, Martikainen PJ, Vasander H 2005: Carbon fluxes from a tropical peat swamp forest floor. *Global Change Biology*, **11**, 1788-1797.
- Jauhiainen J, Vasander H, Rieley J, Page SE 2010: Tropical Peat Carbon Gas Interaction: Technical report 2. Leicester, United Kingdom, University of Leicester.
- Jenkinson DS 2001: The impact of humans on the nitrogen cycle, with focus on temperate arable agriculture. *Plant and Soil*, **228**, 3-15.
- Jia ZJ, Cai ZC, Xu H, Li XP 2001: Effect of rice plants on CH₄ production, transport, oxidation and emission in rice paddy soil. *Plant and Soil*, **230**, 211-221.

- Joabsson A, Christensen TR 2001: Methane emissions from wetlands and their relationship with vascular plants: an Arctic example. *Global Change Biology*, **7**, 919-932.
- Joabsson A, Christensen TR, Wallen B 1999: Vascular plant controls on methane emissions from northern peat forming wetlands. *Tree*, **14**, 385-388.
- Johansson C, Rondhe H, Sanhueza E 1988: Emission of NO in a tropical savanna and a cloud forest during the dry. *Journal of Geophysical Research*, **93**, 7180-7192.
- Jones JB, Mullholland PJ 1998: Methane input and evasion in a hardwood forest stream: effects of subsurface flow from shallow and deep pathways, *Limnology and Oceanography*, **43**, 1243-1250.
- Jungkunst HF, Flessa H, Scherber C, Fiedler S 2008: Groundwater level controls CO₂, N₂O and CH₄ fluxes of three different hydromorphic soil types of a temperate forest ecosystem. *Soil Biology and Biochemistry*, **40**, 2047-2054.
- Junk WJ 2002: Long-term environmental trends and the future of tropical wetlands. *Environmental Conservation*, **29**, 414-435.
- Kalbitz K, Solinger S, Park JH, Michalzik B, Matzner E 2000: Controls on the dynamics of dissolved organic matter in soils: a review. *Soil Science*, **165**, 277-304.
- Kasimir-Klemetsson A, Klemetsson L, Berglund K, Martikainen J, Silvola J, Oenema O 1997: Greenhouse gas emission from farmed organic soils: a review. *Soil Use and Management*, **13**, 245-250.
- Keeney DR, Fillery IR, Marx GP 1979: Effect of temperature on the gaseous nitrogen products of denitrification in a silt loam soil. *Soil Science Society of America Journal*, **43**, 1124-1128.
- Kelker D, Chanton J 1997: The effect of clipping on methane emissions from *Carex*. *Biogeochemistry*, **39**, 37-44.
- Keller M, Goreau TJ, Wofsy SC, Kaplan WA, McElroy MB 1983: Production of nitrous oxide and consumption of methane by forest soils. *Geophysical Research Letter*, **10**, 1156-1159.
- Keller M, Kaplan WA, Wofsy SC, Da Costa JM 1988: Emissions of N₂O from tropical forest soils: response to fertilization with NH₄⁺, NO₃⁻, and PO₄³⁻. *Journal of Geophysical Research*, **93**, 1600-1604.
- Keller M, Reiners WA 1994: Soil atmosphere exchange of N₂O, NO and CH₄ under secondary succession of pasture to forest in the Atlantic lowlands of Costa Rica. *Global Biogeochemical Cycles*, **8**, 399-409.
- Keller M, Veldkamp E, Weitz AM, Reiners WA 1993: Effect of pasture age on soil trace-gas emissions from a deforested area of Costa Rica. *Nature*, **365**, 244-246.
- Kelly CA, Chynoweth DP 1980: Comparison of in situ vitro rates of methane release in freshwater sediments. *Applied and Environmental Microbiology*, **40**, 287-293.
- Kettunen A, Kaitala V, Alm J, Silvola J, Nykänen H, Martikainen JP 1996: Cross-correlation analysis of the dynamics of methane emissions from a boreal peatland. *Global Biogeochemical Cycles*, **10**, 457-471.
- Kettunen A, Kaitala V, Lehtinen A, Lohila A, Alm J, Silvola J, Martikainen PJ 1999: Methane production and oxidation potentials in relation to water table fluctuation in two boreal mires. *Soil Biology and Biochemistry*, **31**, 1741-1749.

- Kiese R, Butterbach-Bahl K 2002: N₂O and CO₂ emissions from three different tropical forest sites in the wet tropics of Queensland, Australia. *Soil Biology and Biochemistry*, **34**, 975-987.
- Kim J, Verma SB 1992: Soil surface CO₂ flux in a Minnesota peatland. *Biogeochemistry*, **18**, 37-51.
- King GM, Roslev P, Skovgaard H 1990: Distribution and rate of methane oxidation in sediments of the Florida everglades. *Applied and Environmental Microbiology*, **56**, 2902-2911.
- King GM, Schnell S 1994: Effect of increasing atmospheric methane concentration on ammonium inhibition of soil methane consumption. *Nature*, **370**, 282-284.
- Kirschbaum MUF 1995: The temperature dependence of soil organic matter decomposition and the effect of global warming on soil organic C storage. *Soil Biology and Biochemistry*, **27**, 753-760.
- Kliewer BA, Gilliam JW 1995: Water table management effects on denitrification and nitrous oxide evolution. *Soil Science Society of America Journal*, **59**, 1694-1701.
- Kling GW, Kipphut GW, Miller MC 1991: Arctic lakes and streams as gas conduits to the atmosphere: Implications for tundra carbon budgets. *Science*, **251**, 298-301.
- Kluge B, Wessolek G, Facklam M, Lorenz M, Schwärzel 2008: Long-term carbon loss and CO₂-C release of drained peatland soils in northeast Germany. *European Journal of Soil Science*, **59**, 1076-1086.
- Knowles R 1985: Microbial transformations as sources and sinks for nitrogen oxides. In: Caldwell DE, Brierley JA, Brierley CL (eds) Planetary ecology. pp. 411-426. Van Nostrand Reinhold. New York, USA.
- Knowles R 1993: Methane: processes of production and consumption. In: Harper LA, Mosier AR, Duxbury JM, Rolston DE (eds) Agricultural ecosystem effects of trace gases and global climate change. pp. 145-165. ASA special publication **55**. American Society of Agronomy. Madison, USA.
- Koschorreck M 2000: Methane turnover in exposed sediments of an Amazon floodplain lake. *Biogeochemistry*, **50**, 195-206.
- Koskinen WC, Keeney DR 1982: Effect of pH on the rate of gaseous products of denitrification in a silt loam soil. *Soil Science Society of America Journal*, **46**, 1165-1167.
- Kozlowski TT 1997: Responses of woody plants to flooding and salinity. *Tree Physiology On-Line Monograph*, **1**, 1-29.
- Krauss KW, Whitbeck JL 2012: Soil greenhouse gas fluxes during wetland forest retreat along the lower Savannah River, Georgia (USA). *Wetlands*, **32**, 73-81.
- Kroeze C, Mosier A, Bouwman L 1999: Closing the global N₂O budget: a retrospective analysis 1500-1994. *Global Biogeochemical Cycles*, **13**, 1-8.
- La Scala N, Marques J, Pereira GT, Corá J.E. 2000: Carbon dioxide emission related to chemical properties of a tropical bare soil. *Soil Biology and Biochemistry*, **32**, 1469-1473.

- Laanbroek HJ 2010: Methane emission from natural wetlands: Interplay between emergent macrophytes and soil microbial processes. A mini-review. *Annals of Botany*, **105**, 141-153.
- Laiho R 2006: Decomposition in peatlands: reconciling seemingly contrasting results on the impacts of lowered water table levels. *Soil Biology and Biochemistry*, **38**, 2011-2024.
- Laiho R, Vasander H, Penttilä T, Laine J 2003: Dynamics of plant-mediated organic matter and nutrient cycling following water-level drawdown in boreal peatlands. *Global Biogeochemical and Cycles*, **17**, 1053.
- Laine A, Sottocornola M, Kiely G, Byrne KA, Wilson D, Tuittila ES 2006: Estimating net ecosystem exchange in a patterned ecosystem: example from blanket bog. *Agricultural and Forest Meteorology*, **138**, 231-243.
- Laine J, Päivänen J 1992: Carbon balance of peatlands and global climatic change: summary. The research programme on climate change. Progress report. Publications of the academy of Finland, Vol. **3/92**, pp. 189-192.
- Lal R, Griffin M, Apt J, Lave L, Morgan MG 2004: Managing soil carbon. *Science*, **304**, 393.
- Lashof DA, Ahuja DR 1990: Relative contributions of greenhouse gas emission to global warming. *Nature*, **344**, 529-531.
- Lazaroff C 2001: Biodiversity gives carbon sinks a boost. *Environment News Service*, April 13.
- Lemke RL, Izaurralde RC, Malhi SS, Arshad MA, Nyborg M 1998: Nitrous oxide emissions from agricultural soils of the boreal and parkland regions of Alberta. *Soil Science Society of America Journal*, **62**, 1096-1102.
- Lessard R, Rochette P, Gregorich EG, Pattey E, Desjardins RL 1996: Nitrous oxide fluxes from manure-amended soil under maize. *Journal of Environmental Quality*, **25**, 1371-1377.
- Liebner S, Schwarzenbach SP, Zeyer J 2012: Methane emissions from an alpine fen in central Switzerland. *Biogeochemistry*, **109**, 287-299.
- Lieffers VJ 1988: Sphagnum and cellulose decomposition in drained and natural areas of an Alberta peatland. *Canadian Journal of Soil Science*, **68**, 755-761.
- Linn DM, Doran 1984: Effect of water-filled pore space on carbon dioxide and nitrous oxide production in tilled and non-tilled soils. *Soil Science Society of America Journal*, **48**, 1267-1272.
- Liptzin D, Silver WL, Detto M 2011: Temporal dynamics in soil oxygen and greenhouse gases in two humid tropical forests. *Ecosystems*, **14**, 171-182.
- Lohila A, Aurela M, Hatakka J, Pihlatie M, Minkkinen K, Penttilä T, Laurila T 2010: Responses of N₂O fluxes to temperature, water table and N deposition in a northern boreal fen. *European Journal of Soil Science*, **61**, 651-661.
- Lohila A, Aurela M, Regina K, Laurila T 2003: Soil and total ecosystem respiration in agricultural fields: effect of soil and crop type. *Plant and Soil*, **251**, 303-317.
- Longdoz B, Yernaux M, Aubinet M 2000: Soil CO₂ efflux measurements in a mixed forest: impact of chamber disturbances, spatial variability and seasonal evolution. *Global Change Biology*, **6**, 907-917.

- Luo J, Tillman RW, White RE, Ball PR 1998: Variation in denitrification activity with soil depth under pasture. *Soil Biology and Biochemistry*, **30**, 897-903.
- Lytle DE, Cronan CS 1998: Comparative soil CO₂ evolution, litter decay, and root dynamics in clear cut and uncut spruce-fir forest. *Forest Ecology and Management*, **103**, 121-128.
- Maas A 1996: A note on the formation of peat deposits in Indonesia. In: Maltby E, Immirzi CP, Safford RJ (eds) Tropical lowland peatlands of Southeast Asia. Proceedings of a workshop on integrated planning and management of tropical lowland peatlands. IUCN, Gland, Switzerland.
- MacDonald JA, Skiba U, Sheppard LJ, Ball B, Roberts JD, Smith KA, Fowler D 1997: The effect of nitrogen deposition and seasonal variability on methane oxidation and nitrous oxide emission rates in an upland spruce plantation and moorland. *Atmospheric Environment*, **31**, 3693-3706.
- MacDonald NW, Zak DR, Pregitzer KS 1995: Temperature effects on kinetics of microbial respiration and net nitrogen and sulfur mineralization. *Soil Science Society of America Journal*, **59**, 233-240.
- Machefert SE, Dise NB, Goulding KWT, Whitehead PG 2004: Nitrous oxide emissions from two riparian ecosystems: key controlling variables. *Water, Air, and Soil Pollution: Focus*, **4**, 427-436.
- Maier CA, Kress LW 2000: Soil CO₂ evolution and root respiration in 11-year old loblolly pine (*Pinus taeda*) plantations as affected by moisture and nutrient availability. *Canadian Journal of Forest Research*, **30**, 347-359.
- Malhi SS, McGill WB, Nyborg M 1990: Nitrate losses in soils: effect of temperature, moisture and substrate concentration. *Soil Biology and Biochemistry*, **22**, 733-737.
- Maljanen M, Liikanen A, Silvola J, Martikainen PJ 2003a: Methane fluxes on agriculture and forested boreal organic soils. *Soil Use and Management*, **19**, 73-79.
- Maljanen M, Liikanen A, Silvola J, Martikainen PJ 2003b: Nitrous oxide emissions from boreal organic soil under different land-use. *Soil Biology and Biochemistry*, **35**, 1-12.
- Mander Ü, Järvijärveoja, Maddison M, Soosaar K, Aavola R, Ostonen I, Salm JO 2012: Reed canary grass cultivation mitigates greenhouse gas emissions from abandoned peat extraction areas. *Global Change Biology Bioenergy*, **4**, 462-474.
- Martikainen P J, Nykänen H, Crill P, Silvola J 1992: The effect of changing water table on methane fluxes at two Finnish mire sites. *Suo*, **43** (4-5), 237-240.
- Martikainen PJ, Nykänen H, Alm J, Silvola J 1995: Change in fluxes of carbon dioxide, methane and nitrous oxide due to forest drainage of mire sites of different trophy. *Plant and Soil*, **168-169**, 571-577.
- Matson PA, McDowell WH, Townsend AR, Vitousek PM 1999: The globalization of N deposition: ecosystem consequences in tropical environments. *Biogeochemistry*, **46**, 67-83.
- Matson PA, Vitousek PM 1990: Ecosystem approach to a global nitrous oxide budget. *BioScience*, **40**, 667-672.
- Matthews E, Fung I 1987: Methane emission from natural wetlands: global distribution, area and environmental characteristics of sources. *Global Biogeochemical Cycles*, **1**, 61-86.

- McHale PJ, Mitchell MJ, Bowles FP 1998: Soil warming in a northern hardwood forest: trace gas fluxes and leaf litter decomposition. *Canadian Journal of Forest Research*, **28**, 1365-1372.
- Megonigal JP, Day FP 1992: Effects of flooding on root and shoot production of bald cypress in large experimental enclosures. *Ecology*, **7**, 1182-1193.
- Megonigal JP, Patrick WH, Faulkner SP 1993: Wetland identification in seasonally flooded forest soil: soil morphology and redox dynamics. *Soil Science Society of America Journal*, **57**, 140-149.
- Melillo JM, Steudler PA, Feigl BJ, Neill C, Garcia D, Piccolo MC, Cerri CC, Tian H 2001: Nitrous oxide emissions from forests and pastures of various ages in the Brazilian Amazon, *Journal of Geophysical Research*, **106**, 34179-34 188.
- Melling L, Goh KJ, Beauvais C, Hatano R 2007: Carbon flow and budget in a young mature oil palm agroecosystem on deep tropical peat. In: Rieley JO, Banks CJ, Radjagukguk B (ed) Proceedings of the international symposium and workshop on tropical peatland. Yogyakarta, Indonesia.
- Melling L, Hatano R, Goh KJ 2005a: Soil CO₂ flux from three ecosystems in tropical peatland of Sarawak, Malaysia. *Tellus*, **57B**, 1-11.
- Melling L, Hatano R, Goh KJ 2005b: Methane fluxes from three ecosystems in tropical peatland of Sarawak, Malaysia. *Soil Biology and Biochemistry*, **37**, 1445-1453.
- Melling L, Hatano R, Goh KJ 2007: Nitrous oxide emissions from three ecosystems in tropical peatland of Sarawak, Malaysia. *Soil Science and Plant Nutrition*, **53**, 792-805.
- Menyailo OV, Huwe B 1999: Activity of denitrification and dynamics of N₂O release in soils under six tree species and grassland in central Siberia. *Journal of Plant Nutrition and Soil Science*, **162**, 533-538.
- Miao Y, Song C, Sun L, Wang X, Meng H, Mao R 2012: Growing season methane emission from a boreal peatland in the continuous permafrost zone of northeast China: effects of active layer depth and vegetation. *Biogeosciences*, **9**, 4455-4464.
- Minami K, Fukushi S 1984: Methods for measuring N₂O flux from water surface and N₂O dissolved in water from agricultural land. *Soil Science and Plant Nutrition*, **30**, 495-502.
- Minami K, Oshawa A, 1990: Emission of nitrous oxide dissolved in drainage water. In: Bouwman AF (ed) Soil and the greenhouse gas effect. John Wiley & Sons Inc. New York.
- Minkinen K, Laine J 2006: Vegetation heterogeneity and ditches create spatial variability in methane fluxes from peatlands drained for forestry. *Plant and Soil*, **285**, 289-304.
- Minkinen K, Laine J, Nykänen H, Martikainen PJ 1997: Importance of drainage ditches in emissions of methane from mires drained for forestry. *Canadian Journal of Forest Research*, **27**, 949-952.
- Mommer L, Pedersen O, Visser EJW 2004: Acclimation of a terrestrial plant to submergence facilities gas exchange under water. *Plant, Cell and Environment*, **27**, 1281-1287.
- Moore S, Gauci V, Evans CD, Page SE 2011: Fluvial organic carbon losses from a Bornean blackwater river, *Biogeosciences*, **8**, 901-909.
- Moore TR 1994: Trace gas emissions from Canadian peatlands and the effect of climatic change. *Wetlands*, **14**, 223-228.

- Moore TR, Dalva M 1993: The influence of temperature and water table position on carbon dioxide and methane emissions from laboratory columns of peatland soils. *Journal of Soil Science*, **44**, 651–664.
- Moore TR, Knowles R 1989: The influence of water table levels on methane and carbon dioxide emissions from peatland soils. *Canadian Journal of Soil Science*, **69**, 33-38.
- Moore TR, Knowles R 1990a: Methane emissions from fen, bog and swamp peatlands in Quebec. *Biogeochemistry*, **11**, 45-61.
- Moore TR, Roulet N, Knowles R 1990b: Spatial and temporal variations of methane flux from subarctic/northern boreal fens. *Global Biogeochemical Cycle*, **4**, 29-46.
- Moore TR, Roulet N 1995: Methane emissions from Canadian peatlands. In: Lal R, Kimble J, Levine E, Steward BA (eds) *Soils and global change*. pp. 153-164. Lewis publishers, Boca Raton, FL.
- Moore TR, Roulet NT, Waddington JM 1998: Uncertainty in predicting the effect of climatic change on the carbon cycling of Canadian peatlands. *Climatic Change*, **40**, 229-245.
- Morishita T, Hatano R, Desyatkin RV 2003: CH₄ flux in an Alas ecosystem formed by forest disturbance near Yakutsk, eastern Siberia, Russia. *Soil Science and Plant Nutrition*, **49**, 369-377.
- Mosier A, Kroeze C, Nevison C, Oenema O, Seitzinger S, van Cleemput O 1998: Closing the global N₂O budget: nitrous oxide emissions through the agricultural nitrogen cycle. *Nutrient Cycling in Agroecosystems*, **53**, 225-248.
- Mosier AR 1989: Chamber and isotope techniques. In: Andreae MO, Schimel DS (eds) *Exchange of trace gases between terrestrial ecosystems and the atmosphere*. pp. 175-187. John Wiley & Sons, New York.
- Mosier AR 1998: Soil processes and global change. *Biology and Fertility of Soils*, **27**, 221-229.
- Mosier AR 2001: Exchange of gaseous nitrogen compounds between agricultural systems and the atmosphere. *Plant and Soil*, **228**, 17-27.
- Mosier AR, Delgado JA 1997: Methane and nitrous oxide fluxes in grasslands in western Puerto Rico. *Chemosphere*, **35**, 2059-2082.
- Mosier AR, Duxbury JM, Freney JR, Heinemeyer O, Minami K 1996: Nitrous oxide emissions from agricultural fields: assessment, measurement and mitigation. In: Cleemput VO, Hofman G, Vermoesen A (eds) *Progress in nitrogen cycling studies*. pp 589-602. Kluwer academic publishers.
- Mosier AR, Zhaoliang Z 2000: Changes in patterns of fertilizer nitrogen use in Asia and its consequences for N₂O emissions from agricultural systems. *Nutrient Cycling in Agroecosystems*, **57**, 107-117.
- Muhanmad NZ, Rieley JO 2002: Management of tropical peatlands in Indonesia: Mega reclamation project in Central Kalimantan. In: Rieley JO, Page SE (eds) *Peatlands for people, natural resources function and sustainable management*. Proceedings of the international symposium on tropical peatlands. August 22-23, 2001. Jakarta, Indonesia. pp. 155–160. BPPT and Indonesian Peat Association.
- Muller C, Sherlock RR, Williams PH 1997: Mechanistic model for nitrous oxide emission via nitrification and denitrification. *Biology and Fertility of Soils*, **24**, 231-238.

- Mummey DL, Smith JL, Bolton H Jr 1994: Nitrous oxide flux from a shrub-steppe ecosystem: sources and regulation. *Soil Biology and Biochemistry*, **26**, 279-286.
- Murdiyarto D, Hergoualc'h K, Verchot LV 2010: Opportunities for reducing greenhouse gas emissions in tropical peatlands, *Proceedings of the National Academy of Science of the United States of America (PNAS)*, **107**, 19655-19660.
- Nakane K, Kohno T, Horikoshi T 1996: Root respiration rate before and just after clear felling in a mature, deciduous, broad-leaved forest. *Ecological Research*, **11**, 111-119.
- Nakane K, Yamamoto M, Tsubota H 1983: Estimation of root respiration rate in a mature forest ecosystem. *Japanese Journal of Ecology*, **33**, 397-408.
- Nedwell DB, Blackburn TH, Wiebe WJ 1994: Dynamic nature and turnover of organic carbon, nitrogen and sulphur in the sediments of a Jamaican mangrove forest. *Marine Ecology Progress Series*, **110**, 223-231.
- Nedwell DB, Watson A 1995: CH₄ production, oxidation and emission in a U.K. ombrotrophic peat bog: Influence of SO₄²⁻ from acid rain. *Soil Biology and Biochemistry*, **27**, 893-903.
- Neuzil SG 1997: Onset and rate of peat and carbon accumulation in four domed ombrogenous peat deposits, Indonesia. In: Rieley JO, Page SE (ed) Tropical peatlands. pp. 55-72. Hereford, United Kingdom. Samara publishing limited.
- Nilsson M, Bohlin E 1993: Methane and carbon dioxide concentrations in bogs and fens-with special reference to the effects of the botanical composition of the peat. *Journal of Ecology*, **81**, 615-625.
- Notohadiprawiro T 1998: Conflict between problem-solving and optimising approach to land resources development policies-the case of Central Kalimantan wetlands. In: Strack M (ed) Peatlands and climate change. 223 pp. International Peat Society (IPS). Jyväskylä, Finland.
- Nouchi I and Mariko S 1993: Mechanism of methane transport by rice plants. In: Oremland RS (ed) Biogeochemistry of global change. Chapman & Hall, Inc.
- Nykänen H, Alm J, Läng K, Silvola T, Martikainen PJ 1995: Emissions of CH₄, N₂O, and CO₂ from a virgin fen drained for grassland in Finland. *Journal of Biogeography*, **22**, 351-357.
- Nykänen H, Alm J, Silvola J, Tolonen K, Martikainen PJ 1998: Methane fluxes on boreal peatlands of different fertility and the effect of long-term experimental lowering of the water table on flux rates. *Global Biogeochemical Cycles*, **12**, 53-69.
- O'Neill BC 2000: The jury is still out on global warming potentials. *Climatic Change*, **44**, 427-443.
- Oberthür S, Ott HE 1999: The Kyoto Protocol: International Climate Policy for the 21st Century. Springer-Verlag Berlin Heidelberg, New York, 172 pp.
- Ojima DS, Valentine DW, Mosier AR, Parton WJ, Schimel DS 1993: Effect of land-use change on methane oxidation in temperate forest and grassland soils. *Chemosphere*, **26**, 675-685.
- Oke TR (ed) 1987: Boundary layer climates 2nd edition. 435 pp. Taylor and Francis group.
- Öquist M, Sundh I 1998: Effects of a transient oxic period on mineralization of organic matter to CH₄ and CO₂ in anoxic peat incubations. *Geomicrobiology Journal*, **15**, 325-333.

- Page S, Hoscilo A, Wösten H, Jauhiainen J, Silvius M, Rieley J, Ritzema H, Tansey K, Graham L, Vasander H, Limin S 2009: Restoration ecology of lowland tropical peatlands in Southeast Asia: current knowledge and future research directions. *Ecosystem*, **12**, 888-905.
- Page SE, Rieley JO, Shotyk W, Weiss D 1999: Interdependence of peat and vegetation in a tropical peat swamp forest. *Philosophical Transactions of the Royal Society of London B series*, **354**, 1885–1897.
- Page SE, Rieley JO, Banks CJ 2011a: Global and regional importance of the tropical peatland carbon pool. *Global Change Biology*, **17**, 798-818.
- Page SE, Morrison R, Malins C, Hooijer A, Rieley JO, Jauhiainen J 2011b: Review of peat surface greenhouse gas emissions from oil palm plantations in Southeast Asia. ICCT White Paper **15**, Washington.
- Page SE, Rieley JO 1998: Tropical peatlands: a review of their natural resource functions, with particular reference to Southeast Asia. *International Peat Journal*, **8**, 95–106.
- Page SE, Siegert F, Rieley JO, Boehm H-DV, Jaya A, Limin SH 2002: The amount of carbon released from peat and forest fires in Indonesia during 1997. *Nature*, **420**, 61–65.
- Page SE, Wüst RAJ, Weiss D, Rieley JO, Shotyk W, Limin SH 2004: A record of late Pleistocene and Holocene carbon accumulation and climate change from an equatorial peat bog (Kalimantan, Indonesia): implications for past, present and future carbon dynamics. *Journal of Quaternary Science*, **19**, 625–635.
- Pangala SR, Moore S, Hornibrook ERC, Gauci V 2013: Trees are major conduits for methane egress from tropical forested wetlands. *New Phytologist*, **197**, 524-531.
- Papen H, Butterbach-Bahl K 1999: A 3-year continuous record of nitrogen trace gas fluxes from untreated and limed soil of a N-saturated spruce and beech forest ecosystem in Germany 1. N₂O emissions. *Journal of Geophysical Research*, **104**, 18487-18503.
- Parkin TB 1987: Soil microsites as a source of denitrification variability. *Soil Science Society of America Journal*, **51**, 1194-1199.
- Parkin TB, Meisinger JJ 1989: Denitrification below the crop rooting zone as influenced by surface tillage. *Journal of Environmental Quality*, **18**, 12-16.
- Parkin TB, Sexstone AJ, Tiedje JM 1985: Adaptation of denitrifying populations to low soil pH. *Applied and Environmental Microbiology*, **49**, 1053-1056.
- Patrick WH, Reddy KR 1976: Nitrification-denitrification reactions in flooded soils and water bottoms: dependence on oxygen supply and ammonium diffusion. *Journal of Environmental Quality*, **5**, 469-472.
- Payne WJ 1981: Denitrification. 214 pp. John Wiley & Sons Inc. New York, USA.
- Pearson J, Woodal J, Clough ECM, Nielsen KH, Schochjorring JK 2002: Production and consumption of NH₄⁺ and NH₃⁻ in trees. In: Gasche R, Papen H, Rennenberg H (eds) Trace gas exchange in forest ecosystems. pp. 53-77. Kluwer academic publishers.

- Pelletier L, Moore TR, Roulet NT, Garneau M, Beaulieu-Audy V 2007: Methane fluxes from three peatlands in the La Grande Rivière watershed, James Bay lowland, Canada. *Journal of Geophysical Research*. **112**. G01018.
- Penman J, Kruger D, Galbally I, Hiraishi T, Nyenzi B, Emmanuel S 2000: Good practice guidance and uncertainty management in national greenhouse gas inventories. Institute for Global Environment Strategies, Japan (for IPCC).
- Pennock DJ, van Kessel C, Farrell RE, Sutherland RA 1992: Landscape-scale variations in denitrification. *Soil Science Society of America Journal*, **56**, 770-776.
- Pete Smith 2004: Carbon sequestration in croplands: the potential in Europe and the global context. *European Journal of Agronomy*, **20**, 229-236.
- Pett-Ridge J, Firestone MK 2005: Redox fluctuation structures microbial communities in a wetland tropical soil. *Applied and Environmental Microbiology*, **71**, 6998-7007.
- Pezeshki SR 2011: Wetland plant responses to soil flooding. *Environmental and Experimental Botany*, **46**, 299-312.
- Pietikäinen J, Vaijärvi E, Ilvesniemi H, Fritze H, Westman CJ 1999: Carbon storage of microbes and roots and the flux of CO₂ across a moisture gradient. *Canadian Journal of Forest Research*, **29**, 1197-1203.
- Poth M, Focht D 1985: 15N kinetic analysis of N₂O production by *Nitrosomonas europaea* : an examination of nitrifier denitrification. *Applied and Environmental Microbiology*, **49**, 1134-1141.
- Prescott CE, Zabek LM, Staley CL, Kabzems R 2000: Decomposition of broadleaf and needle litter in forests of British Columbia: influences of litter type, forest type, and litter mixtures. *Canadian Journal of Forest Research*, **30**, 1742-1750.
- Priemé A, Christensen S, Dobbie KE, Smith KA 1997: Slow increase in rate of methane oxidation in soils with time following land use change from arable agriculture to woodland. *Soil Biology and Biochemistry*, **29**, 1269-1273.
- Qi Y, Xu M 2001: Separating the effects of moisture and temperature on soil CO₂ efflux in a coniferous forest in the Sierra Nevada mountains. *Plant Soil*, **237**, 15-23.
- Raich JW, Potter CS, Bhagawati D 2002: Interannual variability in global soil respiration, 1980-94. *Global Change Biology*, **8**, 800-812.
- Raich JW, Schlesinger WH 1992: The global carbon dioxide flux in soil respiration and its relationship to vegetation and climate. *Tellus*, **44B**, 81-99.
- Raupach MR, Canadell JG 2010: Carbon and the Anthropocene. *Current Opinion in Environmental Sustainability*, **2**, 2010-218.
- Reay DS, Edwards AC, Smith KA 2004: Determinations of nitrous oxide emission from agricultural drainage waters. *Water, Air, and Soil Pollution*, **4**, 107-115.
- Reddy KR, D'Angelo EM 1994: Soil processes regulating water quality in wetlands. In: Mitsch WJ (ed) *Global wetlands: old world and new*. pp. 309-324. Elsevier Science. Amsterdam.
- Renewable Fuels Agency 2010: Palm oil cultivation in Malaysia. London, United Kingdom: the stationary office.

- Rieley JO, Ahmad-Shah AA 1996a: The vegetation of tropical peat swamp forests. In: Maltby E, Immirzi CP, Safford RJ (eds) Tropical lowland peatlands of Southeast Asia, proceedings of a workshop on integrated planning and management of tropical lowland peatlands held at Cisarua, Indonesia. July 3-8, 1992. pp. 55–73. IUCN, Gland, Switzerland.
- Rieley JO, Ahmad-Shah AA, Brady MA 1996b: The extent and nature of tropical peat swamps. In: Maltby E, Immirzi CP, Safford RJ (eds) Tropical lowland peatlands of Southeast Asia, proceedings of a workshop on integrated planning and management of tropical lowlands peatlands, Cisarua, Indonesia. July 3-8, 1992. pp. 17–53. IUCN, Gland, Switzerland.
- Rieley JO, Page SE (eds) 1997: Biodiversity and sustainability of tropical peatlands. Samara publishing Ltd, Cardigan, UK.
- Rieley JO, Page SE 2005: Wise use of tropical peatlands: focus on Southeast Asia. 266 pp. Alterra, Wageningen university and the EU INCO STRAPEAT and RESTORPEAT Partnerships. Wageningen, the Netherlands.
- Rieley JO, Page SE 2008: Carbon budgets under different land uses on tropical peatland. In: Feehan J, Farrell CA (ed) Proceedings 13th international peat congress. Tullamore, Ireland.
- Riley RH, Vitousek PM 1995: Nutrient dynamics and nitrogen trace gas flux during ecosystem development in Montane rain forest. *Ecology*, **76**, 292-304.
- Risch AC, Frank DA 2006: Carbon dioxide fluxes in a spatially and temporally heterogeneous temperate grassland. *Oecologia*, **147**, 291-302.
- Robertson LA, Cornelisse, De Vos P, Hadioetomo R, Kuenen JG 1989: Aerobic denitrification in various heterotrophic nitrifiers. *Antonie van Leeuwenhoek Journal of Microbiology*, **56**, 289-299.
- Robertson LA, Kuenen JG 1988: Heterotrophic nitrification in *Thiosphaera pantotropa*: oxygen uptake and enzyme studies. *Journal of General Microbiology*, **134**, 857-863.
- Rolston DE, Fried M, Goldhamer DA 1976: Denitrification measured directly from nitrogen and nitrous oxide gas fluxes. *Soil Science Society of America Journal*, **40**, 259-266.
- Ronen D, Magaritz M, Almon E 1988: Contaminated aquifers are a forgotten component of the global N₂O budget. *Nature*, **335**, 57-59.
- Roulet NT 2000: Peatlands, carbon storage, greenhouse gases, and the Kyoto Protocol: prospects and significance for Canada. *Wetlands*, **20**, 605-615.
- Rusch H, Rennenberg H 1998: Black alder (*Alnus glutinosa* (L.) Gaertn.) trees mediate methane and nitrous oxide emission from soil to the atmosphere. *Plant and Soil*, **201**, 1-7.
- Ryden JC 1981: N₂O exchange between a grassland soil and the atmosphere. *Nature*, **292**, 235-237.
- Ryden JC, Lund LJ, Focht DD 1978: Direct in-field measurement of nitrous oxide flux from soils. *Soil Science Society of America Journal*, **42**, 731-737.
- Saari A, Heiskanen J, Martikainen PJ 1998: Effect of the organic horizon on methane oxidation and uptake in soil of boreal Scots pine forest. *FEMS Microbiology Ecology*, **26**, 245-255.

- Sahrawat KL, Keeney DR 1986: Nitrous oxide emission from soils. *Advances in Soil Science*, **4**, 103-148.
- Saleska SR, Harte J, Torn MS 1999: The effect of experimental ecosystem warming on CO₂ fluxes in a Montane meadow. *Global Change Biology*, **5**, 125-141.
- Sallantaus T 1992: Leaching in the material balance of peatlands - Preliminary results. *Suo*, **43**, 253-358.
- Scheer C, Wassmann R, Kienzler K, Ibragimov N, Eschanov R 2008: Nitrous oxide emissions from fertilized, irrigated cotton (*Gossypium hirsutum* L.) in the Aral Sea Basin, Uzbekistan: influence of nitrogen applications and irrigation practices. *Soil Biology and Biochemistry*, **40**, 290-301.
- Schiller CL, Hastie DR 1994: Exchange of nitrous-oxide within the Hudson Bay lowland. *Journal of Geophysical Research Atmospheres*, **99**, 1573-1588.
- Schimel D, Melillo J, Tian H, McGuire AD, Kicklighter D, Kittel T, Rosenbloom N, Running S, Thornton P, Ojima D, Parton W, Kelly R, Skykes M, Neilson R, Rizzo B 2000: Contribution of increasing CO₂ and climate to carbon storage by ecosystems in the United States. *Science*, **287**, 2004-2006.
- Schimel JP, Holland EA, Valentine D 1993: Controls on methane flux from terrestrial ecosystems. In: Harper LA, Mosier AR, Duxbury JM, Rolston DE (eds) *Agricultural ecosystem effects of trace gases and global climate change*. pp. 167-182. ASA special publication **55**. American Society of Agronomy. Madison, USA.
- Schlesinger WH 1997: *Biogeochemistry: an analysis of global change* 2nd edition. Academic press. San Diego.
- Schlesinger WH 1999: Carbon sequestration in soils. *Science*, **284**, 2095.
- Schlesinger WH, Andrews JA 2000: Soil respiration and the global carbon cycle. *Biogeochemistry*, **48**, 7-20.
- Schnabel RR, Stout WL 1994: Denitrification loss from two Pennsylvania floodplain soils. *Journal of Environmental Quality*, **23**, 344-348.
- Schnell S, King GM 1994: Mechanistic analysis of ammonium inhibition of atmospheric methane consumption in forest soils. *Applied and Environmental Microbiology*, **60**, 3514-3521.
- Schnitzer M, Gupta UC 1965: Determination of acidity in soil organic matter. *Soil Science Society of America Journal*, **29**, 274-277.
- Schrier-Uijl AP, Veraart AJ, Leffelaar PA, Berendse F, Veenendaal EM 2011: Release of CO₂ and CH₄ from lakes and drainage ditches in temperate wetlands. *Biogeochemistry*, **102**, 265-279.
- Scott KJ, Kelly CA, Rudd JWN 1999: The importance of floating peat to methane fluxes from flooded peatlands. *Biogeochemistry*, **47**, 187-202.
- Sebacher DI, Harriss RC, Bartlett KB 1985: Methane emissions to the atmosphere through aquatic plants. *Journal of Environmental Quality*, **14**, 40-46
- Sehy U, Ruser R, Munch JC 2003: Nitrous oxide fluxes from maize fields: relationship to yield, site-specific fertilization, and soil conditions. *Agriculture Ecosystems and Environment*, **99**, 97-111.

- Seiler W, Conrad 1981: Field measurements of natural and fertilizer-induced N₂O release rates from soils. *Journal of the Air Pollution Control Association*, **37**, 767-772.
- Serca D, Delmas R, Le Roux X, Parsons DAB, Scholes MC, Abbadie L, Lensi R, Ronce O, Labroue L 1998: Comparison of nitrogen monoxide emissions from several African tropical ecosystems and influence of season and fire. *Global Biogeochemical Cycles*, **12**, 637-651.
- Sexstone AJ, Mains CN 1990: Production of CH₄ and ethylene in organic horizons of spruce forest soils. *Soil Biology and Biochemistry*, **22**, 135-139.
- Shannon RD, White JR 1994: A three-year study of controls on methane emissions from two Michigan peatlands. *Biogeochemistry*, **27**, 35-60.
- Sheppard SK, Beckmann M, Lloyd D 2007: The effect of temperature on methane dynamics in soil and peat cores: calculations from membrane inlet mass spectrometry. *Canadian Journal of Soil Science*, **87**, 11-22.
- Shimada S, Takahashi H, Haraguchi A, Kaneko M 2001: The carbon content characteristics of tropical peats in Central Kalimantan, Indonesia: estimating their spatial variability in density. *Biogeochemistry*, **53**, 249-267.
- Shurpali NJ, Verma SB 1998: Micrometeorological measurements of methane flux in a Minnesota peatland during two growing seasons. *Biogeochemistry*, **40**, 1-15.
- Siegert F, Ruecker G, Hinrichs A, Hoffmann AA 2001: Increased damage from fires in logged forests during droughts caused by El niño. *Nature*, **414**, 437-440.
- Silver WL, Lugo AE, Keller M 1999: Soil oxygen availability and biogeochemistry along rainfall and topographic gradients in upland wet tropical forest soils. *Biogeochemistry*, **44**, 301-328.
- Silvius MJ 1989: Indonesia. In: Scott DA (compiler) A directory of Asian wetlands, pp. 981-1109. IUCN, Gland, Switzerland and Cambridge.
- Silvola J, Alm J, Ahlholm U, Nykänen H, Martikainen PJ 1996: CO₂ fluxes from peat in boreal mires under varying temperature and moisture conditions. *Journal of Ecology*, **84**, 219-228.
- Šimek M, Hopkins DW 1999: Regulation of potential denitrification by soil pH in long-term fertilized arable soils. *Biology and Fertility of Soils*, **30**, 41-47.
- Singh JS, Gupta SR 1977: Plant decomposition and soil respiration in terrestrial ecosystems. *Botanical Reviews*, **43**, 449-528.
- Sitaula BK, Bakken LR 1993: Nitrous oxide release from spruce forest soil: relationships with nitrification, methane uptake, temperature, moisture and fertilization. *Soil Biology and Biochemistry*, **25**, 1415-1421.
- Sitaula BK, Bakken LR, Abrahamsen G 1995: N-fertilization and soil acidification effects on N₂O and CO₂ emission from temperate pine forest soil. *Soil Biology and Biochemistry*, **27**, 1401-1408.
- Skiba U, Fowler D, Smith K 1994: Emissions of NO and N₂O from soils. *Environmental Monitoring and Assessment*, **31**, 153-158.
- Skiba U, Sheppard J, MacDonald J, Fowler D 1998: Some key environmental variables controlling nitrous oxide emissions from agricultural and semi-natural soils in Scotland. *Atmospheric Environment*, **32**, 3311-3320.

- Skiba U, Sheppard LJ, Pitcairn CER, van Dijk S, Rossall MJ 1999: The effect of N deposition on nitrous oxide and nitric oxide emissions from temperate forest soils. *Water, Air, and Soil Pollution*, **116**, 89-98.
- Skiba U, Smith KA 2000: The control of nitrous oxide emissions from agricultural and natural soils. *Chemosphere*, **2**, 379-386.
- Skiba U, Smith KA, Fowler D 1993: Nitrification and denitrification as sources of nitric oxide and nitrous oxide in a sandy loam soil. *Soil Biology and Biochemistry*, **25**, 1527-1536.
- Slemr F, Conrad R, Seiler W 1984: Nitrous oxide emissions from fertilized and unfertilized soils in subtropical region (Andalusia, Spain). *Journal of Atmospheric Chemistry*, **1**, 159-169.
- Slemr F, Seiler W 1984: Field measurements of NO and N₂O emissions from fertilized and unfertilized soils. *Journal of Atmospheric Chemistry*, **2**, 1-24.
- Smith KA, McTaggart IP, Dobbie KE, Conen F 1998: Emissions of N₂O from Scottish agricultural soils, as a function of fertilizer N. *Nutrient Cycling in Agroecosystems*, **52**, 123-130.
- Smith SJ, Wigley TML 2000: Global warming potentials: 1. Climatic implications of emissions reductions. *Climatic Change*, **44**, 445-457.
- Society of Chemical Engineers, Japan 1999: Handbook of chemical engineering (in Japanese) 7thed. Maruzen. Tokyo.
- Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL (eds) Climate change 2007: the physical science basis. Contribution of working group I to the fourth assessment report of the intergovernmental panel on climate change. Cambridge University press, Cambridge. UK.
- Sorensen KW 1993: Indonesian peat swamp forests and their role as a carbon sink. *Chemosphere*, **27**, 1065-1082.
- Stanier RY, Adelberg EA, Ingraham JL 1978: General microbiology 5th edition. Macmillan Press, London.
- Stehfest E, Bouwman L 2006: N₂O and NO emission from agricultural fields and soils under natural vegetation: summarizing available measurement data and modelling of global annual emissions. *Nutrient Cycling in Agroecosystems*, **74**, 207-128.
- Steinkamp R, Butterbach-Bahl K, Papen H 2001: Methane oxidation by soils of an N limited and N fertilized spruce forest in the Black Forest, Germany. *Journal Soil Biology and Biochemistry*, **33**, 145-153.
- Stuedler PA, Bowden RD, Melillo JM, Aber JD 1989: Influence of nitrogen fertilization on methane uptake in temperate forest soils. *Nature*, **341**, 314-316.
- Stuedler PA, Melillo JM, Bowden RD, Castro MS, Lugo AE 1991: The effect of natural and human disturbance on soil nitrogen dynamics and trace gas fluxes in a Puerto Rican wet forest. *Biotropica*, **23**, 356-363.
- Stuedler PA, Melillo JM, Feigl BJ, Neill C, Piccolo MC, Cerri CC 1996: Consequence of forest-to-pasture conversion on CH₄ fluxes in the Brazilian Amazon Basin. *Journal of Geophysical Research*, **10**, 18547-18554.
- Stevens RJ, Laughlin RJ, Malone JP 1998: Soil pH affects the processes reducing nitrate to nitrous oxide and di-nitrogen. *Soil Biology and Biochemistry*, **30**, 1119-1126.

- Strack M (ed) 2008: Peatlands and climate change. International Peat Society, Jyväskylä, Finland.
- Strack M, Waddington JM, Tuittila ES 2004: Effect of water table drawdown on northern peatland methane dynamics: implications for climate change. *Global Biogeochemical Cycles*, **18**, GB4003.
- Striegl RG, McConnaughey TA, Thorstenson DC, Weeks EP, Woodward JC 1992: Consumption of atmospheric methane by desert soils. *Nature*, **357**, 145-147.
- Sulistiyanto Y, Rieley J, Limin S 2004: Nutrient dynamics in different sub-types of ombrotrophic peat swamp forest in Central Kalimantan, Indonesia. In: Pangala RS, Moore S, Hornibrook ERC, Gauci V 2013: Trees are major conduits for methane egress from tropical forested wetlands. *New Phytologist*, **197**, 524-531.
- Sundari S, Hirano T, Yamada H, Kusin K, Limin S 2012: Effect of groundwater level on soil respiration in tropical peat swamp forests. *Journal of Agricultural Meteorology*, **68**, 121-134.
- Sundh I, Mikkilä, Nilsson M, Svensson BH 1995: Potential aerobic methane oxidation in a sphagnum-dominated peatland-controlling factors and relation to methane emission. *Soil Biology and Biochemistry*, **27**, 829-837.
- Suzuki S, Nagano T, Waijaroen S 1999: Influences of deforestation on carbon balance in a natural tropical peat swamp forest in Thailand. *Environmental Control in Biology*, **37**, 115-128.
- Svensson BH, Rosswall T 1984: In situ methane production from acid peat in plant communities with different moisture regimes in a subarctic mire. *Oikos*, **43**, 341-350.
- Takahashi H, Simada S, Ibie B, Usup A, Yudha, Limin SH 2002: Annual changes of water balance and a drought index in a tropical peat swamp forest of Central Kalimantan, Indonesia. In: Rieley JO, Page SE (ed) Proceedings of the international symposium on tropical peatlands. Jakarta, Indonesia.
- Takahashi H, Usup A, Hayasaka H, Limin SH 2003: Estimation of ground water level in a peat swamp forest as an index of peat/forest fire. In: Osaki M, Iwakuma T, Kohyama, *et al* (eds) Proceedings of the international symposium on land management and biodiversity in Southeast Asia. September 17–20, 2002. pp. 311–314. Bali, Indonesia. Hokkaido university and Indonesian institute of sciences, Sapporo, Japan and Bogor, Indonesia.
- Takakai F, Morishita T, Hashidoko Y, Darung U, Kuramochi K, Dohong S, Limin SH, Hatano R 2006: Effects of agricultural land-use change and forest fire on N₂O emission from tropical peatlands, Central Kalimantan, Indonesia. *Soil Science and Plant Nutrition*, **52**, 662–674.
- Terazawa K, Ishizuka S, Sakata T, Yamada K, Takashi M 2007: Methane emissions from stems of *Fraxinus manshurica* var. *japonica* trees in a floodplain forest. *Soil Biology and Biochemistry*, **39**, 2689-2692.
- Terry RE, Tate III RL, Duxbury JM 1981: Nitrous oxide emissions from drained, cultivated organic soils of south Florida. *Journal of the Air Pollution Control Association*, **31**, 1173-1176.

- Thomas GW, Hargrove WL 1984: The chemistry of soil acidity. In: Adams F (ed) Soil acidity and liming. Agronomy monograph 12. 2nd edition. pp. 3-56. American Society of Agronomy Inc, Crop Science Society of America Inc, and Soil Science Society of America Inc. Madison, Wisconsin, USA.
- Thomas KL, Benstead J, Davies KL, Lloyd D 1996: Role of wetland plants in the diurnal control of CH₄ and CO₂ fluxes in peat. *Soil Biology and Biochemistry*, **28**, 17–23.
- Tiedje JM 1988: Ecology of denitrification and dissimilatory nitrate reduction to ammonium. In: Zehnder AJB (ed) Biology of anaerobic microorganisms. pp. 179–244. Wiley-Interscience. Ontario, Canada.
- Toma Y, Hatano R 2007: Effect of crop residue C:N ratio on N₂O emissions from Gray Lowland soil in Mikasa, Hokkaido, Japan. *Soil Science and Plant Nutrition*, **53**, 198-205.
- Toma Y, Takakai F, Darung U, Kuramochi K, Limin SH, Dohong S, Hatano R 2011: Nitrous oxide emission derived from soil organic matter decomposition from tropical agricultural peat soil in Central Kalimantan, Indonesia. *Soil Science and Plant Nutrition*, **57**, 436-451.
- Topp E, Pattey E 1997: Soils as sources and sinks for atmospheric methane. *Canadian Journal of Soil Science*, **77**, 167-178.
- Trujillo-Tapia N, Cruz Mondragón C, Vásquez-Murrieta MS, Van Cleemput O, Dendooven L 2008: Inorganic N dynamics and N₂O production from tannery effluents irrigated soil under different water regimes and fertilizer application rate: A laboratory study. *Applied Soil Ecology*, **38**, 279-288.
- Tuah SJ, Jamal YM, Limin SH 2003: Nutritional characteristics in leaves of plants native to tropical peat swamps and heath forests of Central Kalimantan, Indonesia. *Tropics*, **12**, 221–245.
- Tuah SJ, Osaki M, Matsubara T, Limin SH, Istomo, Segah H, Putir PE 2001: Study on leaf elemental characteristics of native plants grown in a distinct ecosystem of tropical peatland in Central Kalimantan. In: Environmental conservation and land use management of wetland ecosystem in Southeast Asia, Annual report of JSPS-LIPI core university program for April 2000-March 2001. pp. 69-84. Graduate school of environmental earth science, Hokkaido university, Sapporo.
- Ueda S, Go UCS, Yoshioka T, Yoshida N, Wada E, Miyajima T, Sugimoto A, Boontanon N, Vijarnsorn P and Boonprakub S 2000: Dynamics of dissolved O₂, CO₂, CH₄, and N₂O in a tropical coastal swamp in southern Thailand. *Biogeochemistry*, **49**, 191-215.
- Updegraff K, Pastor J, Bridgham SD, Johnston CA 1995: Environmental and substrate controls over carbon and nitrogen mineralization in northern wetlands. *Ecological Applications*, **5**, 151-163.
- Urban NR, Bayley SE, Eisenreich SJ 1989: Export of dissolved organic carbon and acidity from peatlands. *Water Resources Research*, **25**, 1619-1628.
- Usup A, Hashimoto Y, Takahashi H, Hayasaka H 2004: Combustion and thermal characteristics of peat fire in tropical peatland in Central Kalimantan, Indonesia. *Tropics*, **14**, 1-19.

- Valentini R, Matteucci G, Dolman AJ, Schulze ED, Rebmann C, Moors EJ, Granier A, Gross P, Jensen NO, Pilegaard K, Lindroth A, Grelle A, Bernhofer C, Grünwald T, Aubinet M, Ceulemans R, Kowalski AS, Vesala T, Rannik Ü, Berbigier P, Loustau D, Guðmundsson J, Thorgeirsson H, Ibrom A, Morgenstern K, Clement R, Moncrieff J, Montagnani L, Minerbi S, Jarvis PG 2000: Respiration as the main determinant of carbon balance in European forest. *Nature*, **404**, 861-865.
- Van Cleemput O 1998: Subsoils: chemo-and biological denitrification, N₂O and N₂ emissions. *Nutrient Cycling in Agroecosystems*, **52**, 187-194.
- van der Gon HACD, Neue HU 1996: Oxidation of methane in the rhizosphere of rice plants. *Biology and Fertility of Soils*, **22**, 359-366.
- van der Nat FJWA, Middelburg JJ 1998: Seasonal variation in methane oxidation by rhizosphere of *Phragmites australis* and *Scirpus lacustris*. *Aquatic Botany*, **61**, 95-110.
- van der Werf GR, Dempewolf J, Trigg SN, Randerson JT, Kasibhatia PS, Giglio L, Murdiyarso D, Peters W, Morton DC, Collatz GJ, Dolman AJ, DeFries R S 2008: Climate regulations of fire emissions and deforestation in equatorial Asia. *Proceedings of the national academy of sciences of the United States of America*, **105**, 20350-20355.
- Van Kessel C, Pennock DJ, Farrel RE 1993: Seasonal variations in denitrification and nitrous oxide evolution at the landscape scale. *Soil Science Society of America Journal*, **57**, 988-995.
- Van Miegroet H, Cole DW 1985: Acidification sources in red alder and Douglas fir soils-importance of nitrification. *Soil Science Society of America Journal*, **49**, 1274-1279.
- Vann CD, Megonigal JP 2003: Elevated CO₂ and water depth regulation of methane emissions: comparison of woody and non-woody wetland plant species. *Biogeochemistry*, **63**, 117-134.
- Velthof GL, Brader AB, Oenema O 1996: Seasonal variations in nitrous oxide losses from managed grasslands in the Netherlands. *Plant and Soil*, **181**, 263-274.
- Venterea RT, Rolston DE 2000: Mechanism and kinetics of nitric and nitrous oxide production during nitrification in agricultural soil. *Global Change Biology*, **6**, 303-316.
- Verchot LV, Davidson EA, Cattânio JH, Ackerman IL, Erickson HE, Keller M 1999: Land use change and biogeochemical controls of nitrogen oxide emissions from soils in eastern Amazonia. *Global Biogeochemical Cycles*, **13**, 31-46.
- Verville JH, Hobbie SE, Chapin III FS, Hooper DU 1998: Response of tundra CH₄ and CO₂ flux to manipulation of temperature and vegetation. *Biogeochemistry*, **41**, 215-235.
- Verwer C, van der Meer P, Nabuurs GJ 2008: Review of carbon estimates and other greenhouse gas emissions from oil palm cultivation on tropical peatlands – Identifying the gaps in knowledge. Wageningen, Alterra-rapport, **1731**, ISSN 1566-7197.
- Verwer C, van der Meer PJ 2010: Carbon pools in tropical peat forest: towards a reference value for forest biomass in relatively undisturbed peat swamp forest in Southeast Asia. Wageningen, Alterra, Alterra-report, **2108**, ISSN 1566-7197.
- Vijarnsorn P 1996: Peatlands in Southeast Asia: a regional perspective. In: Maltby E, Immirzi CP, Safford RJ (eds) Tropical lowland peatlands of Southeast Asia. Proceedings of a workshop on integrated planning and management of tropical lowland peatlands. IUCN, Gland, Switzerland.

- Vitousek P, Matson P, Volkamann C, Maass JM, Garcia G 1989: Nitrous oxide flux from dry tropical forest. *Global Biogeochemical Cycles*, **3**, 375-382.
- Von Arnold K, Weslien P, Nilsson M, Svensson BH, Klemedtsson 2005: Fluxes of CO₂, CH₄ and N₂O from drained coniferous forests on organic soils. *Forest Ecology and Management*, **210**, 239-254.
- Waddington JM, Roulet NY, Swanson RV 1996: Water table control of CH₄ emission enhancement by vascular plants in boreal peatlands. *Journal of Geophysical Research*, **101**, 22,775-22,785.
- Wang MX, Shangguan XJ 1996: CH₄ emission from various rice fields in P.R. China. *Theoretical and Applied Climatology*, **55**, 129-138.
- Wang ZP, DeLaune RD, Masscheleyn PH, Patrick WH Jr 1993: Soil redox and pH effects on methane production in a flooded rice soil. *Soil Science Society of America Journal*, **57**, 382-385.
- Watson RT, Meira Filho LG, Sanhueza E, Janetos A 1992: Greenhouse gases: sources and sinks. In: Houghton JT, Callander BA, Varney SK (eds) *Climate change 1992: the supplementary report to the IPCC scientific assessment*. Cambridge university press. Cambridge, New York.
- Watson RT, Rodhe H, Oeschger H, Siegenthaler U 1990: Greenhouse gases and aerosols. In: Houghton JT, Jenkins GJ, Ephraums JJ (eds) *Climate change: the IPCC scientific assessment*. Cambridge university press. Cambridge, New York.
- Werner C, Zheng XH, Tang JW, Xie BH, Liu CY, Kiese R, Butterbach-Bahl K 2006: N₂O, CH₄, and CO₂ emissions from seasonal tropical rain forest and a rubber plantation in Southwest China. *Plant Soil*, **289**, 335-353.
- Wetlands International – Indonesia programme 2004: Peta luas sebaran lahan gambut dan kandungan karbon di pulau Kalimantan (Maps of area of peatland distribution and carbon content in Kalimantan. 1st edition. ISBN: 979-95899-9-1.
- Whalen SC 2005: Biogeochemistry of methane exchange between natural wetlands and the atmosphere. *Environmental Engineering Science*, **22**, 73-94.
- Whalen SC, Reeburgh WS 1988: A methane flux time series for tundra environments. *Global Biogeochemistry Cycles*, **2**, 399-409.
- Whalen SC, Reeburgh WS 1996: Moisture and temperature sensitivity of CH₄ oxidation in boreal soils. *Soil Biology and Biochemistry*, **10/11**, 1271-1281.
- Whalen SC, Reeburgh WS 2000: Methane oxidation, production, and emission at contrasting sites in a Boreal Bog. *Geomicrobiology Journal*, **17**, 237-251.
- Whalen SC, Reeburgh WS, Kiser KS 1991: Methane consumption and emission by taiga. *Global Biogeochemical Cycles*, **5**, 261-273
- Whalen SC, Reeburgh WS 1990: A methane flux transect along the trans-Alaska pipeline haul road. *Tellus*, **42B**, 237-249.
- Whiting GJ, Chanton JP 1992: Plant-dependent CH₄ emission in a subarctic Canadian fen. *Global Biogeochemical Cycles*, **6**, 225-231.
- Whiting GJ, Chanton JP 1993: Primary production control of methane emission from wetlands. *Nature*, **364**, 794-795.

- Whiting GJ, Chanton JP 1996: Control of the diurnal pattern of methane emission from emergent aquatic macrophytes by gas transport mechanisms. *Aquatic Botany*, **54**, 237-253.
- Widén B, Majdi H 2001: Soil CO₂ efflux and root respiration at three sites in a mixed pine and spruce forest: Seasonal and diurnal variation. *Canadian Journal of Forest Research*, **31**, 786-796.
- Williams CJ, Yavitt, JB 2010: Temperate wetland methanogenesis: The importance of vegetation type and root ethanol production. *Soil Science Society of America Journal*, **74**: 317-325.
- Wösten JHM, Clymans E, Page S, Rieley J, Limin S 2008: Peat-water interrelationships in a tropical peatland ecosystem in Southeast Asia. *Catena*, **73**, 212-224.
- Wösten JHM, Hooijer A, Siderius C, Rais DS, Idris A, Rieley JO 2006a: Tropical peatland water management modelling of the Air Hitam Laut catchment in Indonesia. *International Journal of River Basin Management*, **4** (4), 233-244.
- Wösten JHM, Van den Berg J, Van Eijk, P, Gevers GJM, Giesen WBJT, Hooijer A, Idris A, Leenman, PH, Rais DS, Siderius C, Silvius MJ, Suryadiputra N, Tricahyo Wibisono I 2006b: Interrelationships between hydrology and ecology in fire degraded tropical peat swamp forests. *International Journal of Water Resources Development*, **22** (1), 157-174.
- Wösten JMH, Ismail AB, van Wijk ALM 1997: Peat subsidence and its practical implications: a case study in Malaysia. *Geoderma*, **78**, 25-36.
- Xu M, Qi Y 2001: Soil-surface CO₂ efflux and its spatial and temporal variations in a young ponderosa pine plantation in northern California. *Global Change Biology*, **7**, 667-677.
- Yagi K, Tsuruta H, Kanda K, Minami K 1996: Effect of water management on methane emission from a Japanese rice paddy field: automated methane monitoring. *Global Biogeochemical Cycles*, **10**, 255-267.
- Yavitt JB, Downey DM, Lancaster E, Lang GE 1990: Methane consumption in decomposing Sphagnum-derived peat. *Soil Biology and Biochemistry*, **22**, 441-447.
- Yoshida T, Alexander M 1970: Nitrous oxide formation by *Nitrosomonas europaea* and heterotrophic microorganisms. *Soil Science Society of America Journal*, **34**, 880-882.
- Yozzo DJ, Titre JP (eds) 1996: Planning and evaluating restoration of aquatic habitats. Final report to US army engineer waterways experiment station. Vicksburg, Mississippi, USA. In: McCormick PV, Laing JA 2003: Effects of increased phosphorus loading on dissolved oxygen in a subtropical wetland, the Florida Everglades. *Wetlands Ecology and Management*, **11**, 199-216.
- Yu K, DeLaune RD 2006: A modified soil diffusion chamber for gas profile analysis. *Soil Science Society of America Journal*, **70**, 1237-1241.
- Yu K, Faulkner SP, Baldwin MJ 2008: Effect of hydrological condition on nitrate oxide, methane, and carbon dioxide dynamics in a bottom land hardwood forest and its implication for soil carbon sequestration. *Global Change Biology*, **14**, 798-812.
- Yu KW, Wang ZP, Vermoesen A, Patrick WHJr, Cleemput O Van 2001: Nitrous oxide and methane emissions from different soil suspensions: effect of soil redox status. *Biology and Fertility of Soils*, **34**, 25-30.

- Zak DR, Holmes WE, MacDonald NW, Pregitzer KS 1999: Soil temperature, matric potential, and the kinetics of microbial respiration and nitrogen mineralization. *Soil Science Society of America Journal*, **63**, 575-584.
- Zech W, Senesi N, Guggenberger G, Kaiser K, Lehmann J, Miano TM, Miltner A, Schroth G 1997: Factors controlling humification and mineralization of soil organic matter in the tropics. *Geoderma*, **79**, 117-161.
- Zhu R, Liu Y, Ma J, Xu H, Sun L 2008: Nitrous oxide flux to the atmosphere from two coastal tundra wetlands in eastern Antarctica. *Atmospheric Environment*, **42**, 2437-2447.
- Zogg GP, Zak DR, Ringerberg DB, White DC, MacDonald NW, Pregitzer KS 1999: Compositional and functional shifts in microbial communities due to soil warming. *Soil Science Society of America Journal*, **61**, 475-481.