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**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)**

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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 $+10 \text{ cm}$). 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

No.	Figure caption	Page
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

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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 (Kozlowski, 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.

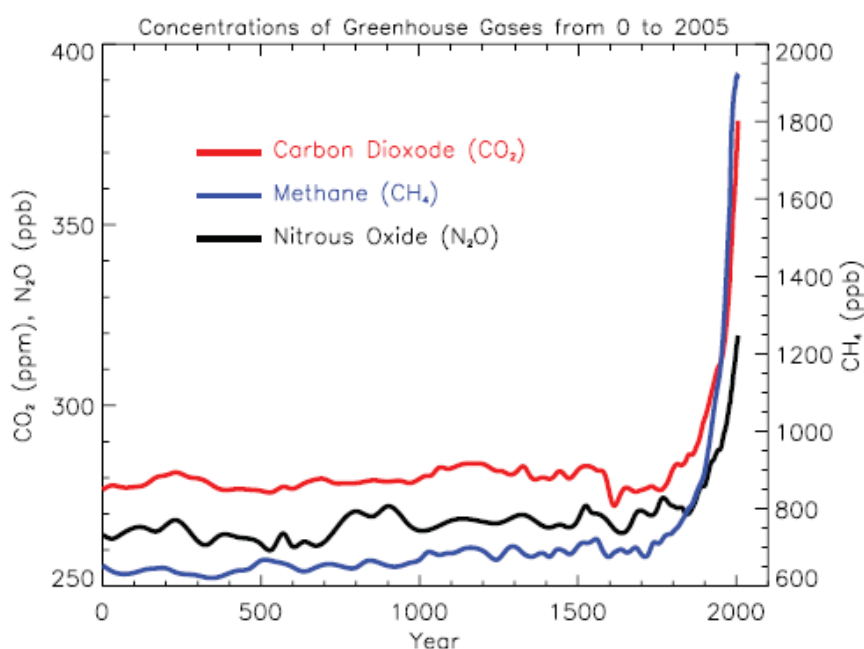


Fig. 2.1 Change in abundance of major GHGs for the last thousand years (IPCC, 2007).

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**).

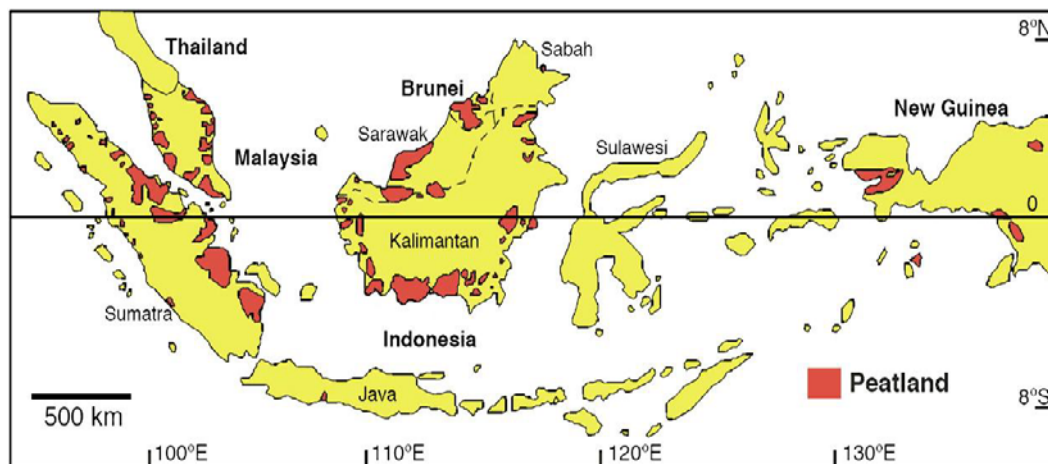


Fig. 2.2 Map of peatland distribution in the Southeast Asia (Page *et al.*, 2004).

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 Andriess, 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 (Andriess, 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-Klemetsson *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).

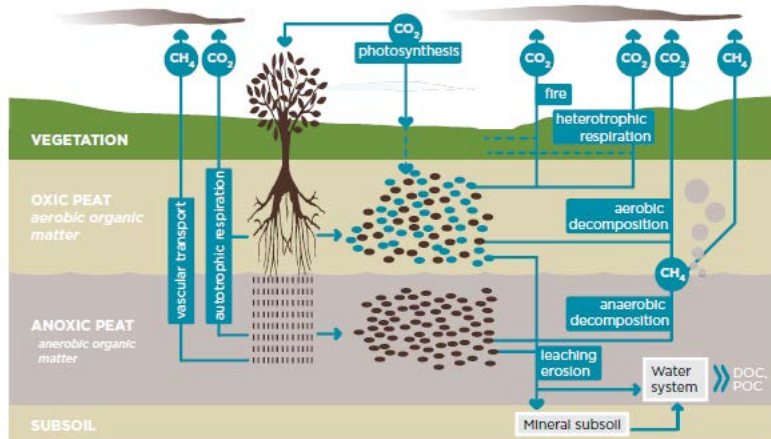


Fig. 2.4 Schematic representation of C cycle processes, flow paths, and stores in tropical peatland (after Jauhiainen *et al.*, 2001).

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-Klemetsson *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).

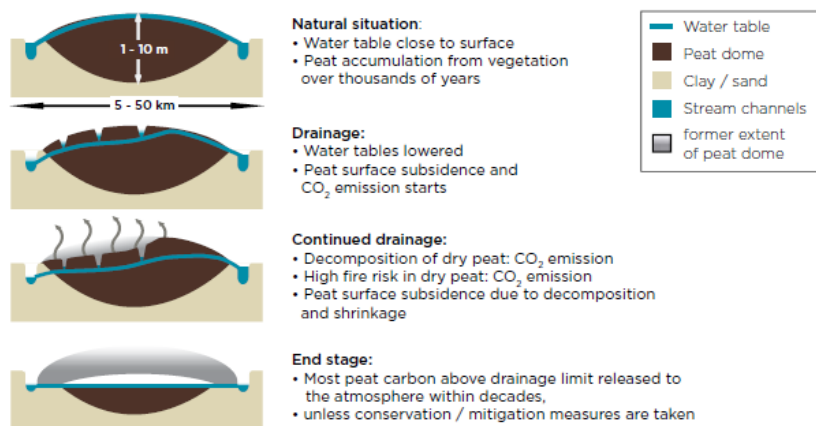


Fig. 2.5 Schematic illustration of drainage effects on a peatland dome (modified from Delft Hydraulics, 2006).

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; Liefvers, 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 Borken, 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 (*Tristaniopsis* 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; Minkinen 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 ($2.15\text{--}2.54\ \mu\text{mol m}^{-2}\text{ s}^{-1}$) than in the cooler Hawaiian montane peatlands ($0.83\text{--}1.81\ \mu\text{mol m}^{-2}\text{ s}^{-1}$), 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 kgCO₂-C m⁻² 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 kgCO₂-C m⁻² 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 g m⁻² yr⁻¹; 166.0 to 169.2 molCO₂ m⁻² yr⁻¹) and deforested site CO₂ fluxes (2781 to 2608 g m⁻² yr⁻¹; 63.2 to 59.3 molCO₂ m⁻² 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 $109.0\pm9.0\ \text{mol m}^{-2}$

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; Sallantauss, 1992).

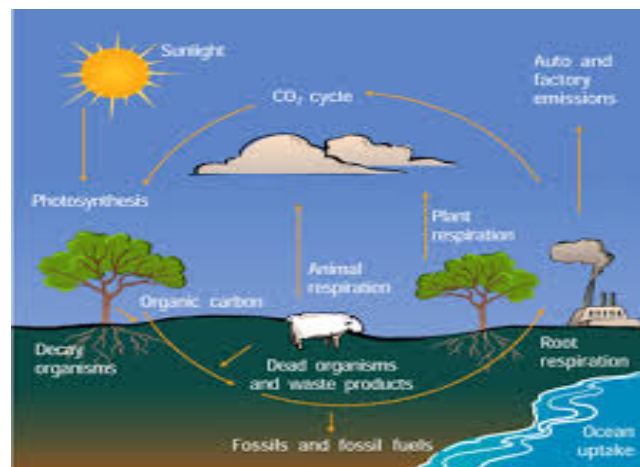


Fig. 2.6 Schemes of C fluxes and factors controlling them in the ecosystems.

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 *et 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; Götter 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 of 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-Klemetsson *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 N_2O 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 CO_2 as the principal C source for biomass formation. N_2O production is enhanced in soils having a high mineralization capacity to form NH_4^+ or treated with nitrifiable forms of N.

In most forest soils, the importance of nitrification or denitrification as the main source of N_2O is not static. It can switch very rapidly, as the soil aeration state within the biologically active site changes due to rainfall or increased O_2 demand caused by the presence of easily mineralizable organic matter. Though denitrification has been demonstrated to be a very important source of N_2O 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 N_2O 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 N_2O in soil seems to be produced during NH_4^+ oxidation and NO_3^- reduction, the exact production mechanism is usually not easily understood for a specific field situation.

Release of N_2O 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 N_2O 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 N_2O 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 NH_4^+ oxidation by nitrifying bacteria and NO_3^- reduction by denitrifying bacteria; and
2. The size of the holes in the pipe through which gaseous N_2O , NO, and N_2 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.

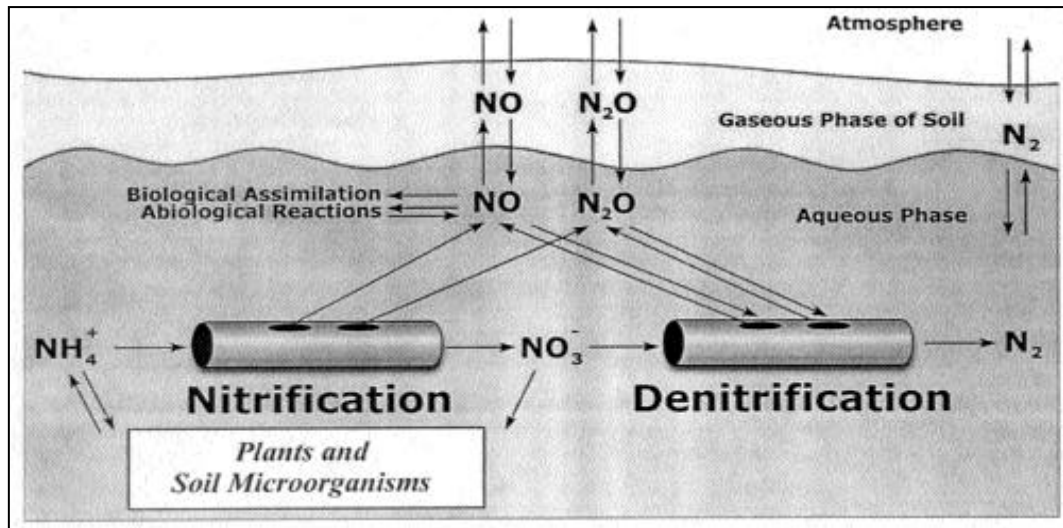


Fig. 2.7 Diagram of the “hole-in-the-pipe” conceptual model describing soil emissions of N_2O and NO (Davidson *et al.*, 2000).

The first level of regulation determines the total amount of N oxides produced ($\text{NO} + \text{N}_2\text{O}$) while the second level of regulation determines the relative importance of NO and N_2O 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: CO_2 , CH_4 and N_2O . GWP values for emissions of CO_2 , CH_4 , and N_2O on a molar basis for a 100-year time horizon are 1:25:298. Although CH_4 and especially N_2O are far lower atmospheric concentrations than CO_2 , 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 CH_4 is large. Microbial processes (methanogenesis and CH_4 oxidation) play a major role in the release of CH_4 to the atmosphere. The high concentrations of CH_4 in shallow soil are particularly common where standing water is present, because it impedes entry of O_2 to support methanotrophy (Bartlett *et al.*, 1988; Koschorreck, 2000). Svensson and Rosswall (1984) reported the variations in dissolved CH_4 profiles in acidic peat soils; CH_4 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 CH_4 (10 cm depth intervals) in bogs, but because of the indirect method employed, high spatial resolution was not achieved. Thus, Sebach *et al.* (1985) observed a peak in the vertical distribution of dissolved CH_4 in sediment pore water near the rhizomes of aquatic plants.

The concentrations of N_2O 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 N_2O can be dissolved in the soil solution. The maximal concentrations of dissolved N_2O reported are up to three orders of magnitude above ambient. On the other hand, temporary N_2O 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 N_2O in soils is very different. In some studies the concentrations of N_2O 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 N_2O concentrations in the soil profiles is in part the result of seasonal and chemical factors regulating N_2O dynamics. Some investigations revealed seasonal dynamics of N_2O 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 N_2O with the soil solution and further out-gassing into the atmosphere. Leaching of dissolved N_2O requires high N_2O production rates and great amounts of infiltrating water causing fast preferential flow (Dowdell *et al.*, 1979; Schnabel and Stout, 1994). In the relation of N_2O production, the denitrification being a major N_2O production process, clear relationships between increasing concentrations of N_2O and decreasing availability of O_2 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						Mean 1.4 to 5.8 (2002)		Strack <i>et al.</i> , 2004.
										Mean 0.3 to 0.7 (2002)		
3	Peatland in the Mer Bleue, near Ottawa in eastern Ontario, Canada	Drained site						0.1-2.5	0-0.6		Blodau <i>et al.</i> , 2007.	
4	The Mer Bleue site near Ottawa, eastern Ontario, Canada	<i>Vegetations of mosses</i>	Oligotrophic peatland	June 2003 to December 2004								
5	The Mer Bleue site near Ottawa, eastern Ontario, Canada	Mosses and shrubs	Oligotrophic peatland						0.01-0.190		Blodau and Moore, 2003.	
6	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.	
7	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.	
	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							Ding <i>et al.</i> , 2004.	
					21.6-66.5				24-229			
									8-202			
					17.2- 45.0							
					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 Gleysol 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 1.2-39.3		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> <i>Open ditch</i>	Podzols, brown forest soil, and noncalcareous gleys soil	August 29, 2002		Mean 28.2 Mean 8.5 Mean 3.8	Reay <i>et al.</i> , 2004.
13	The Pärnu Lowland in western part of Estonia 58° 34' 20" N, 24° 23' 15" E	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 Hollows Hummocks	Typic tropofibrists	in 1999, 2000, and 2001 during wet and dry season	Mean 139 Mean 484 Maximum 0.35	Jauhainen <i>et al.</i> , 2005.
19	Tropical peatland, upper Sebangau River Southeast of Palangka Raya, Central Kalimantan	Forest Hollows Hummocks 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

		<i>Plot B</i>		<i>Plot C</i>	
			Mean 154 (dry season) and 637 (wet season)		
			Mean 94 (dry season) and 2441 (wet season)		
22	Peatland at Marchantaria Island, an island in Amazon River, near Manaus, Brazil 03° 13.94' S, 59° 56.765' W	Shallow part of lake Camaleão (Floodplain ecosystem)	October to November 1996 and 1997	0.175-1.380	Koscho-reck, 2000.
23	The Daeng and Bacho swamp peatland near Narathiwat province in Thailand	Mangrove forest	Wet season (Nov-Dec, 1991-1993) and dry season (Jul. 1992 and Aug. 1993)	0-3.290	Ueda <i>et al.</i> , 2000.
		Transition of the sandy plains and the swamps		0.00001-0.417	
23	Peatland in southernmost Patagonia, Tierra del Fuego 54° 45' S, 68° 20' W	Vascular plants	Bog peatland	Mean < 0.001 (rooting zone)	Fritz <i>et al.</i> , 2011.
			Spring 2006 to summer 2007	Mean 0.166 (> 170 cm depth)	
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	Marsh (Wetland ecosystem)	-20.6-85.6	Zhu <i>et al.</i> , 2008.
		<i>Lynobya</i> , <i>Nostoc</i> , <i>Cosmarium</i> , etc	December 2005 to February 2006		

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 (Muhanmad 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).

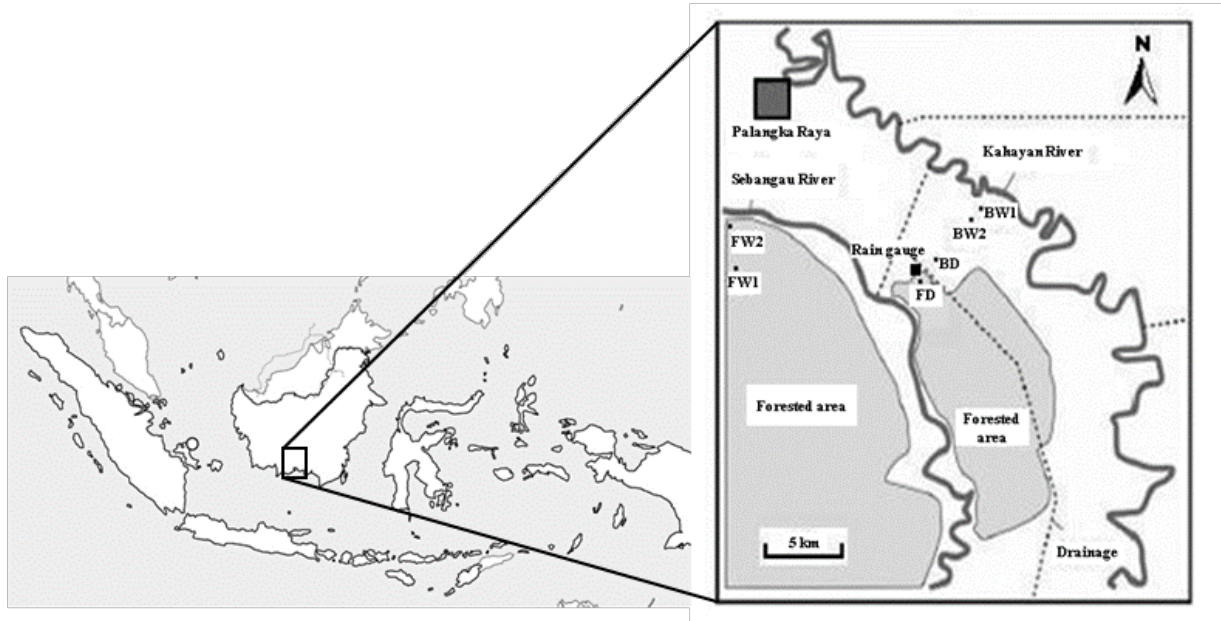


Fig. 3.1 Locations of field observation sites.

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 (*Stenochlaenapalustris*) 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.

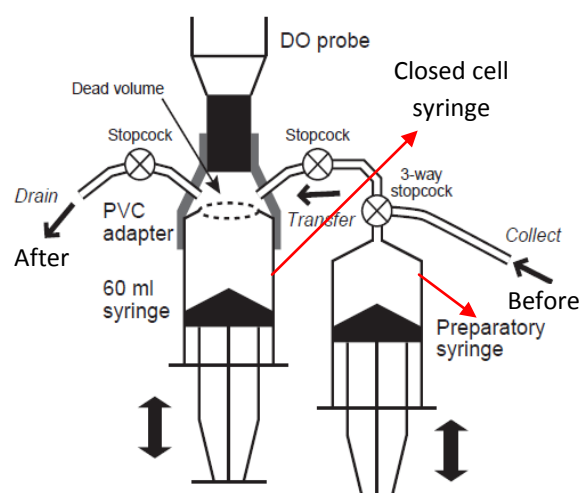


Fig. 4.3 Diagram of an in-situ DO measurement.

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.

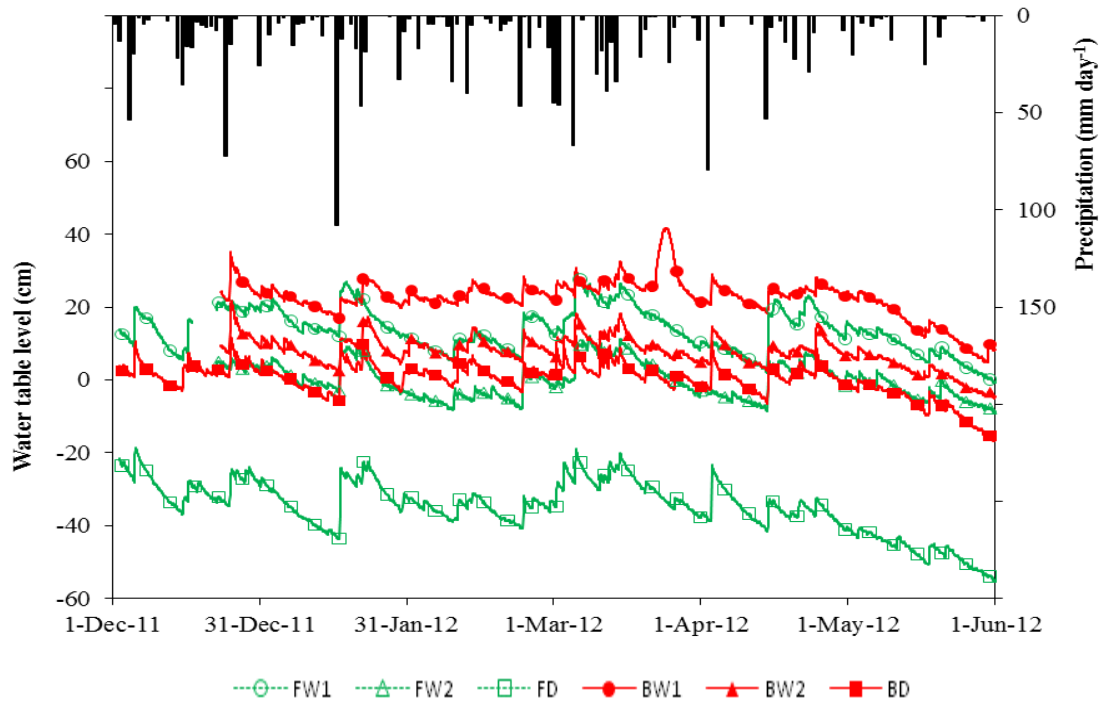


Fig. 4.5 Seasonal changes in precipitation (bar) and water table level at each site (line with symbol).

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.

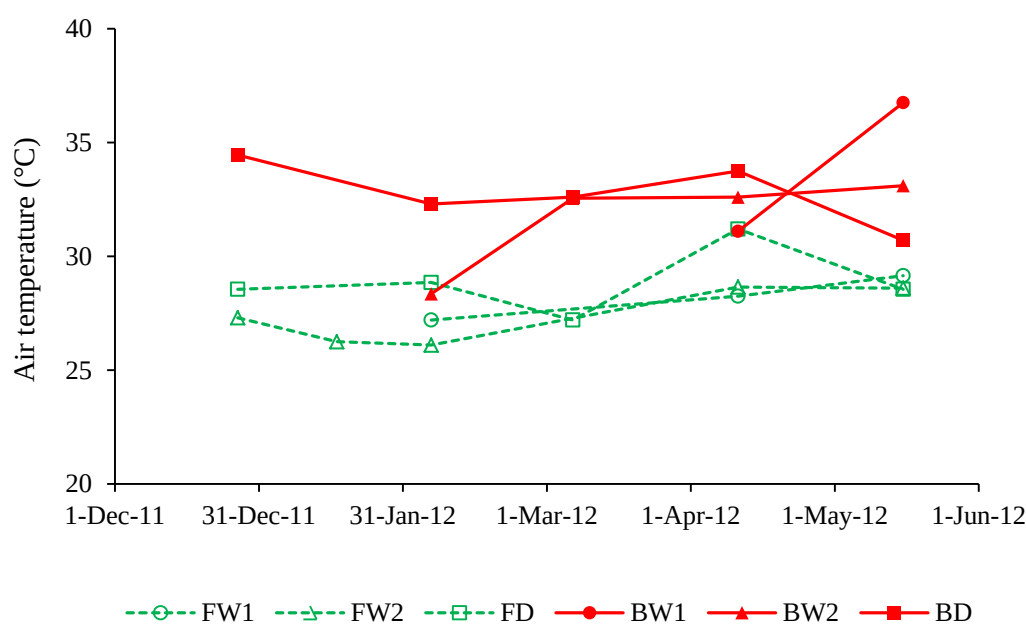


Fig. 4.6 Seasonal variations in air temperature at each study sites.

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.

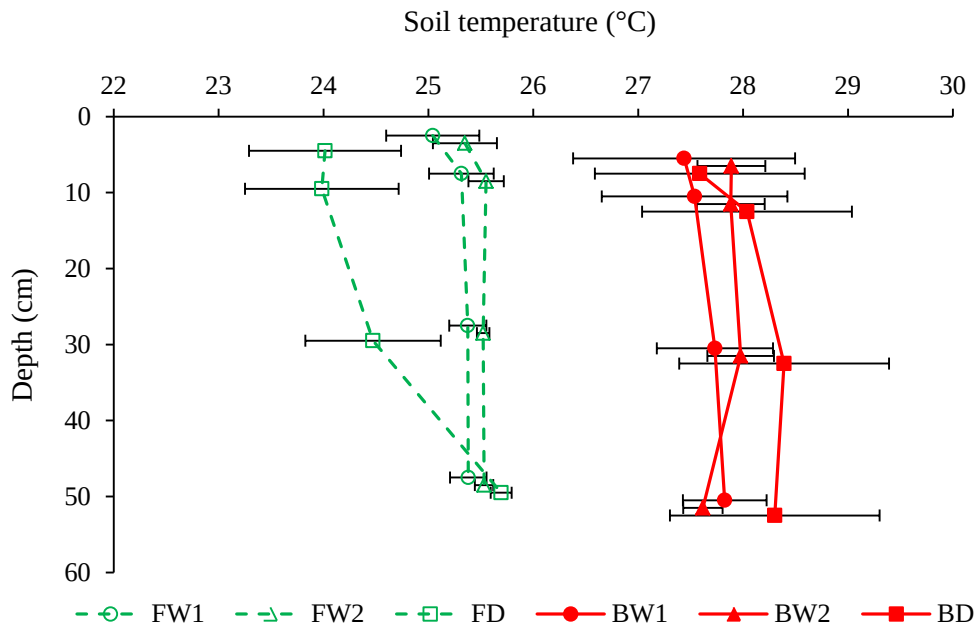


Fig. 4.7 Vertical distributions in soil temperature at all soil depths. Error bars show standard deviation. For better identifiability, the depths in the profiles were slightly shifted from their actual depths (5, 10, 30, and 50 cm).

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).

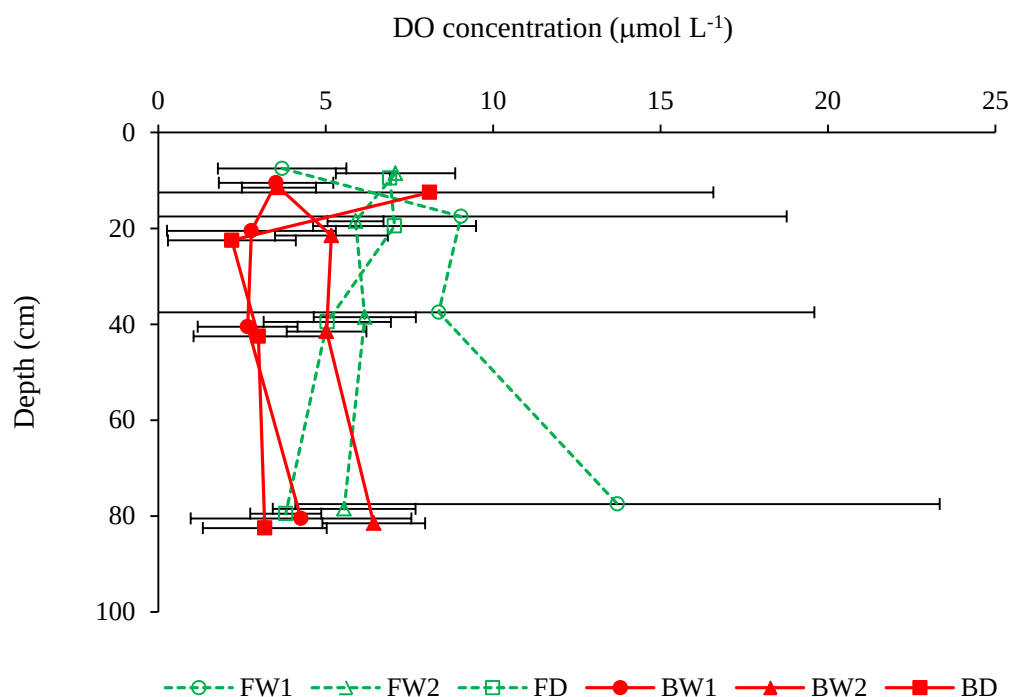


Fig. 4.8 Vertical distributions of DO concentration in subsurface pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation. For better identifiability, the depths in the profiles were slightly shifted from their actual depths (10, 20, 40, and 80 cm).

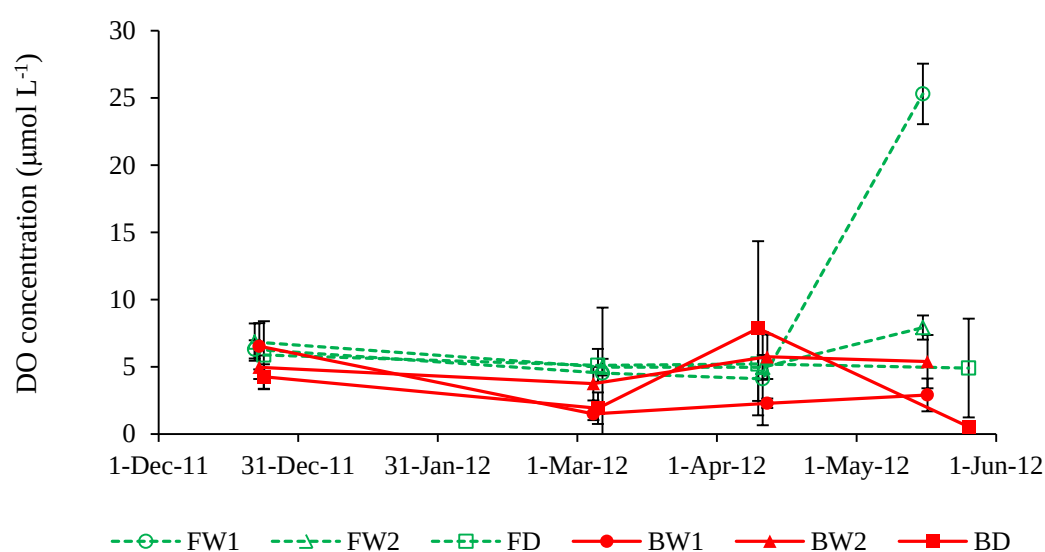


Fig. 4.9 Seasonal variations of DO concentration in subsurface pore water averaged for depths of 10–80 cm at each site. Error bars show standard deviation.

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).

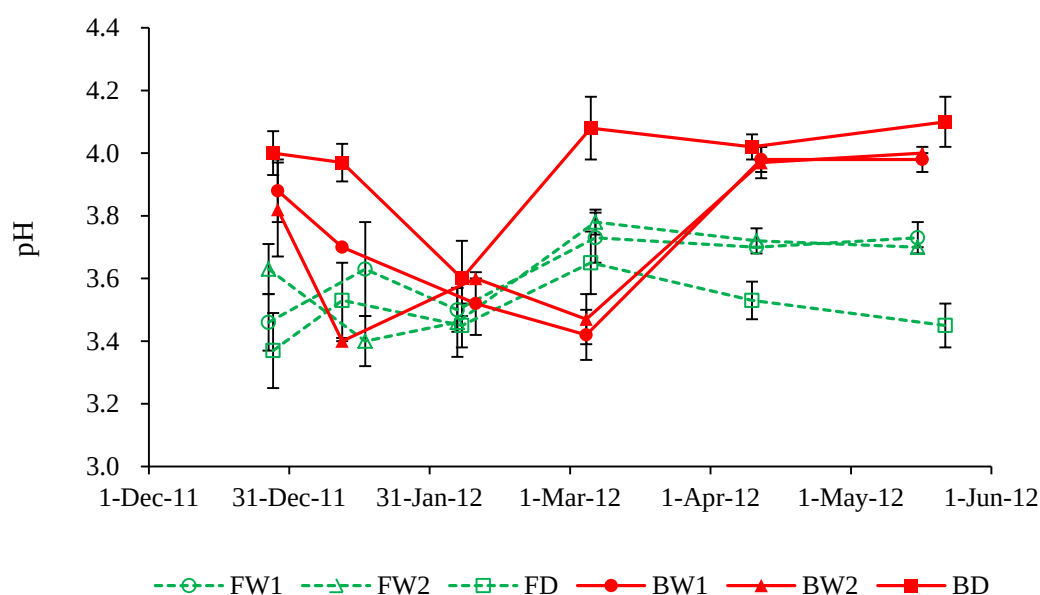


Fig. 4.10 Seasonal variations of pH in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation.

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.

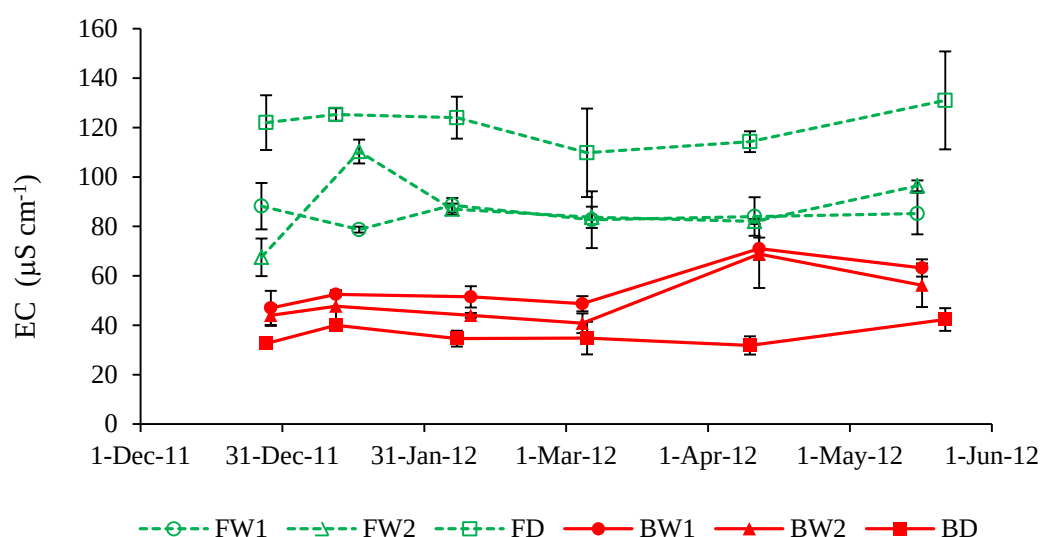


Fig. 4.11 Seasonal variations of EC in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation.

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 $\text{BW2} > \text{BD} > \text{FD} > \text{FW1} > \text{FW2}$ (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).

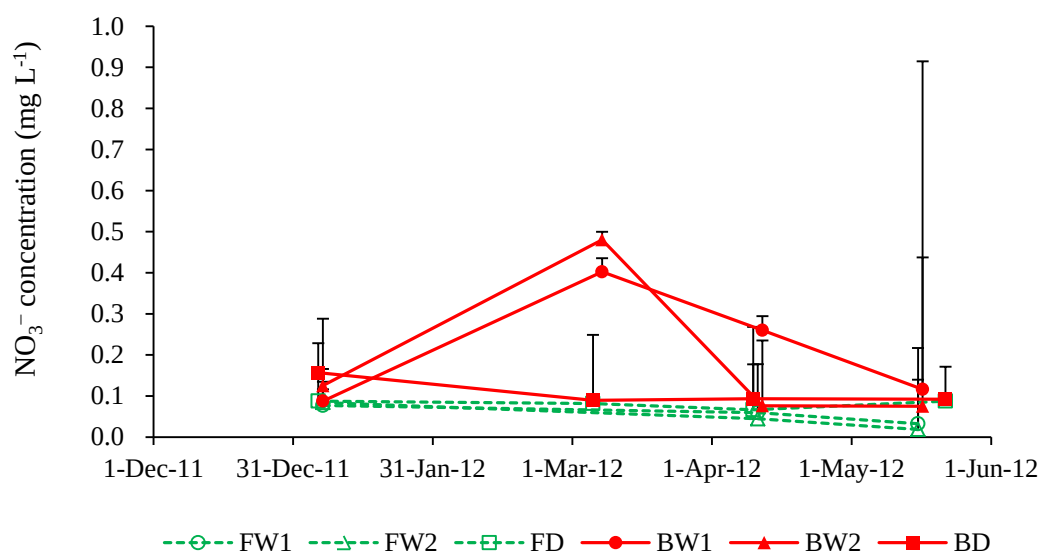


Fig. 4.12 Seasonal variations of NO_3^- concentration in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation (positive side only).

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**).

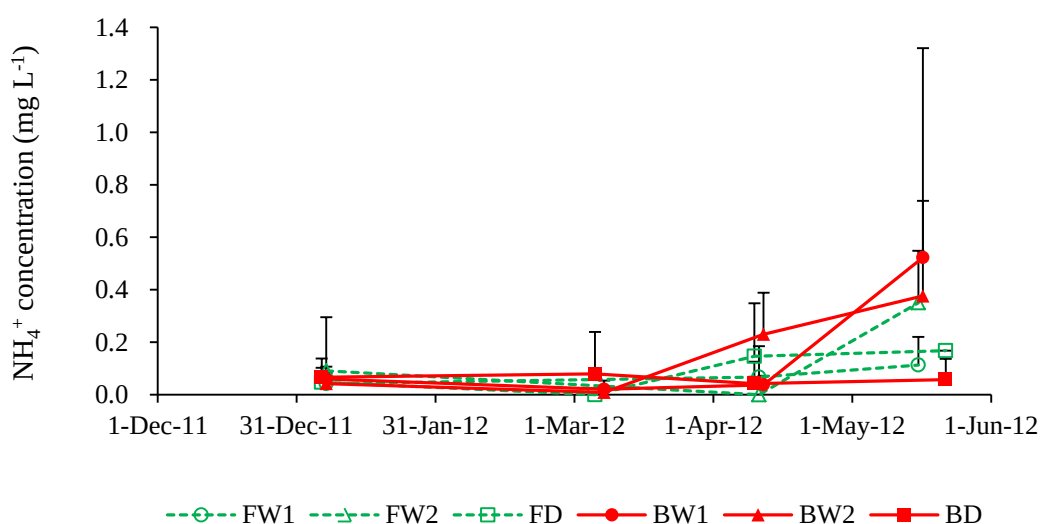


Fig. 4.13 Seasonal variations of NH_4^+ concentration in peat water averaged for depths of 10–80 cm at each site. Error bars show standard deviation (positive side only).

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.

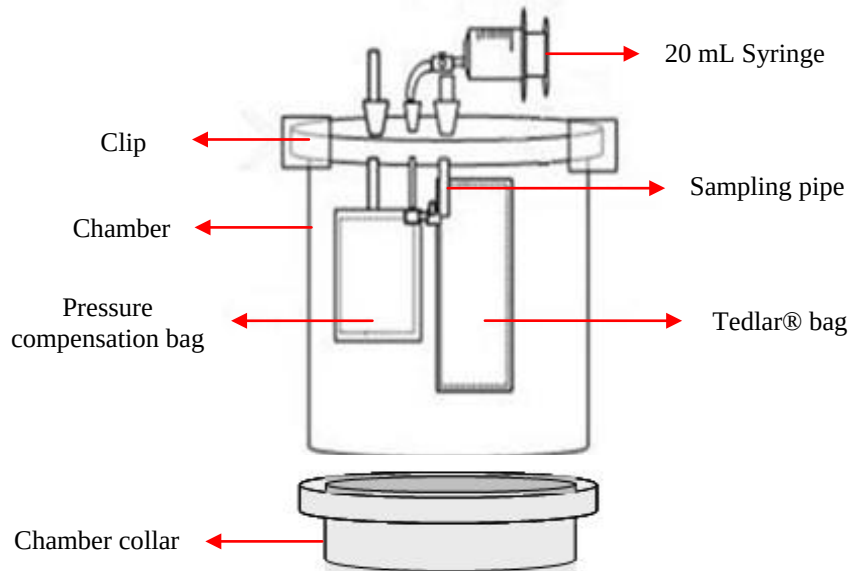


Fig. 5.1 Schematic diagram of a closed chamber (Toma and Hatano, 2007).

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 ($5.75 \pm 6.66 \text{ mgC m}^{-2} \text{ h}^{-1}$) 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 ($0.0084 \pm 0.0321 \text{ mgC m}^{-2} \text{ h}^{-1}$). The site-specific averages of CH₄ fluxes ranged from 0.0084 to $5.75 \text{ mgC m}^{-2} \text{ h}^{-1}$ (**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 significant higher than under drained conditions in both land use during the field observation.

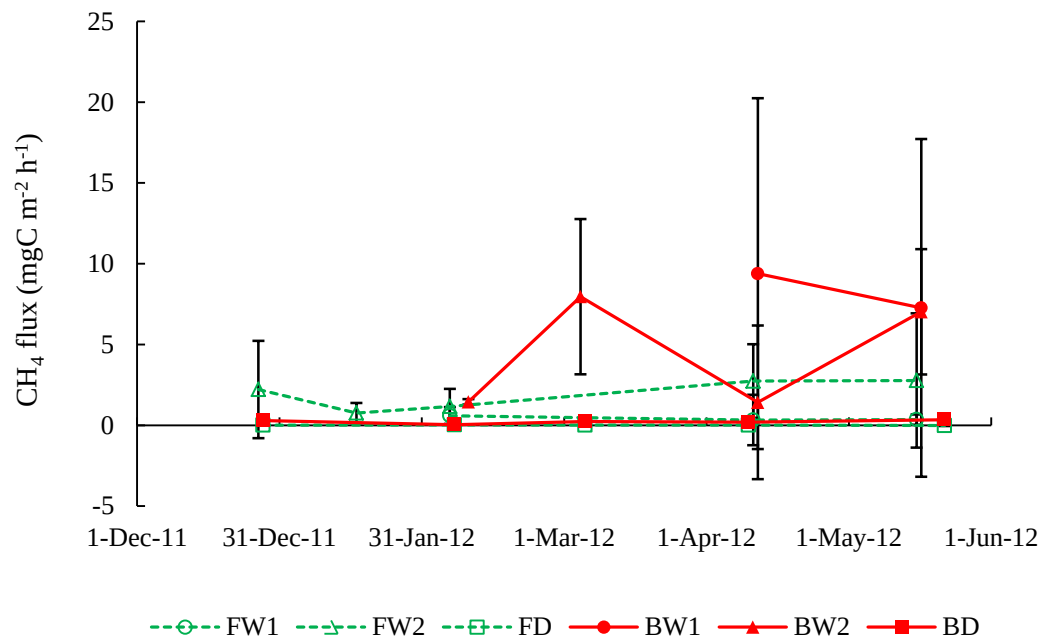


Fig. 5.2 Seasonal variations of CH₄ fluxes measure at the ground surface in the study sites. Error bars show standard deviation.

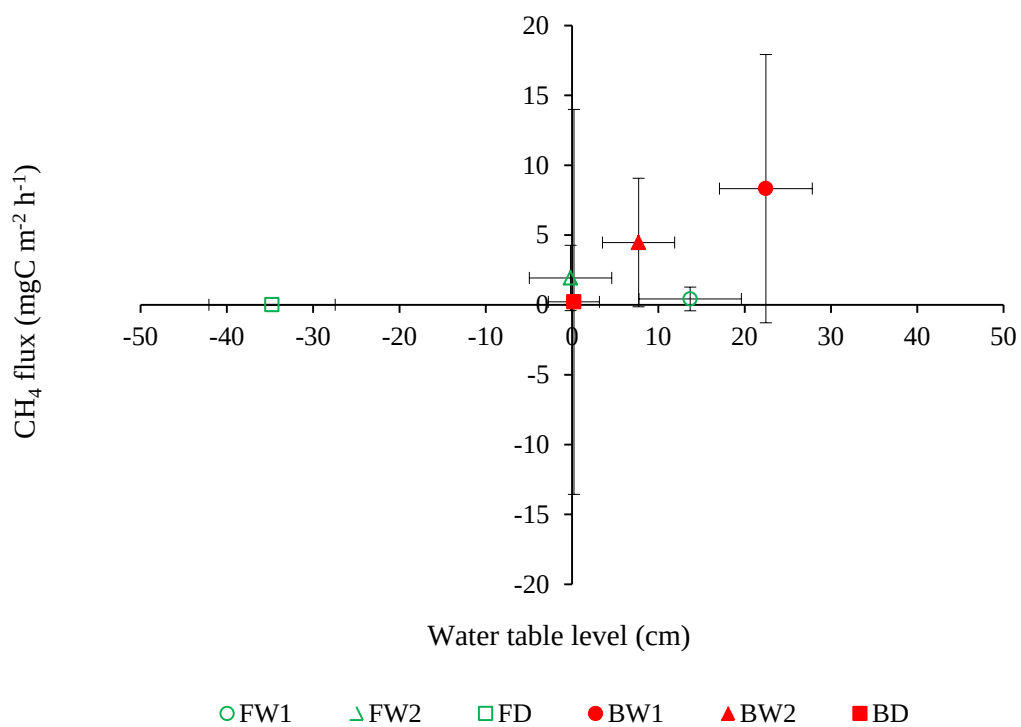


Fig. 5.3 Relationship between CH₄ flux and water table level averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

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**).

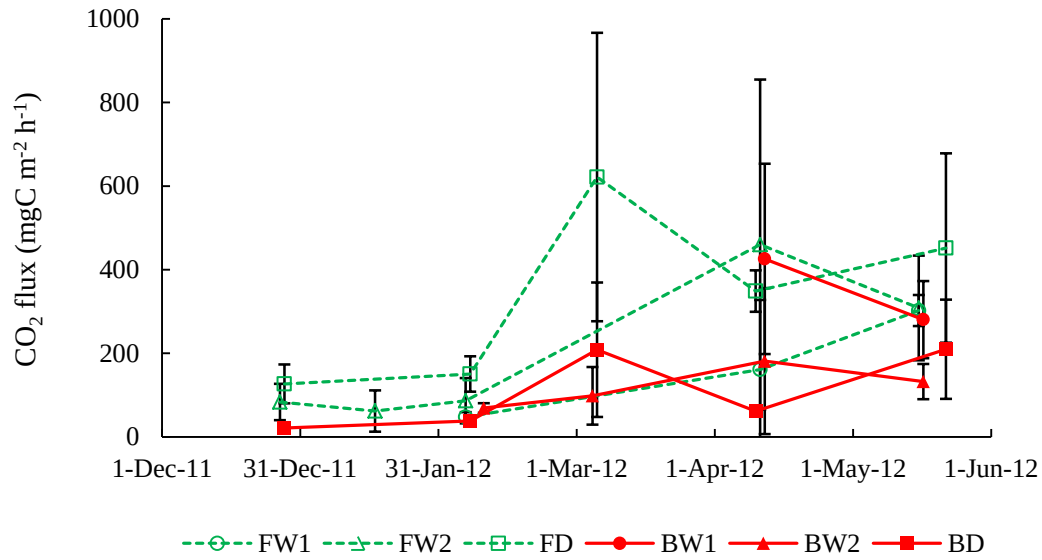


Fig. 5.4 Seasonal variations of CO₂ fluxes measure at the ground surface in the study sites. Error bars show standard deviation.

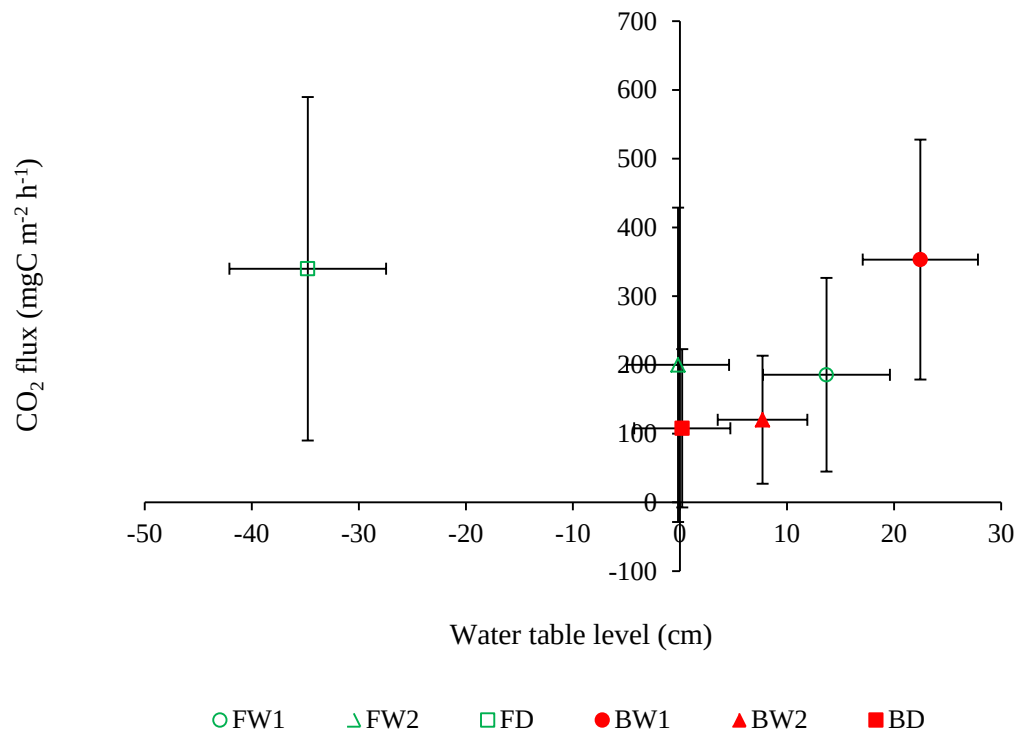


Fig. 5.5 Relationship between CO₂ flux and water table level averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

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 $\mu\text{gN m}^{-2} \text{h}^{-1}$.

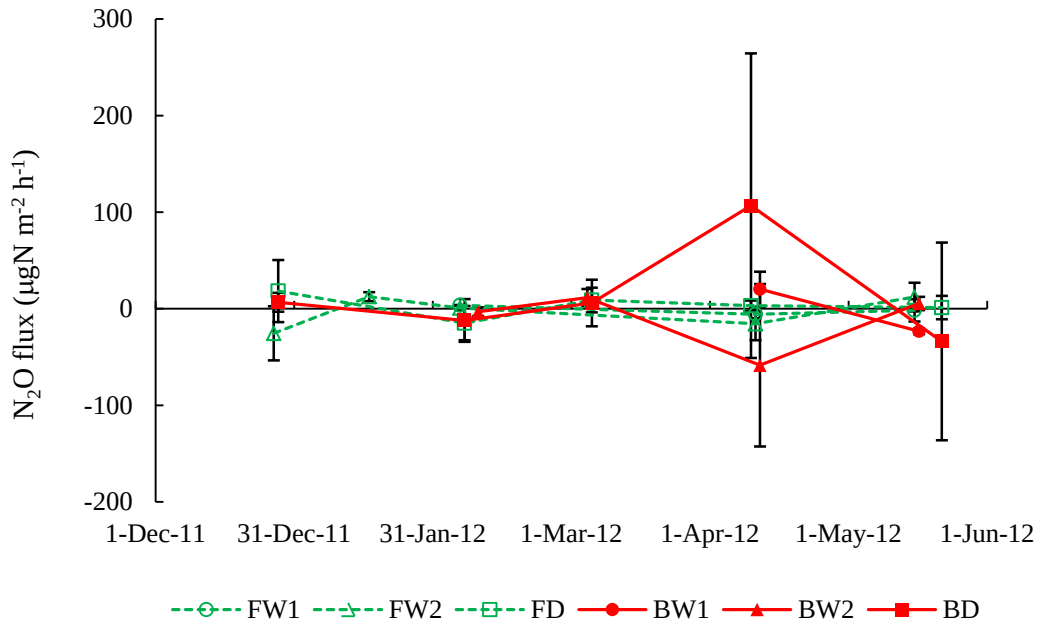


Fig. 5.6 Seasonal variations of N₂O fluxes measure at the ground surface in the study sites. Error bars show standard deviation.

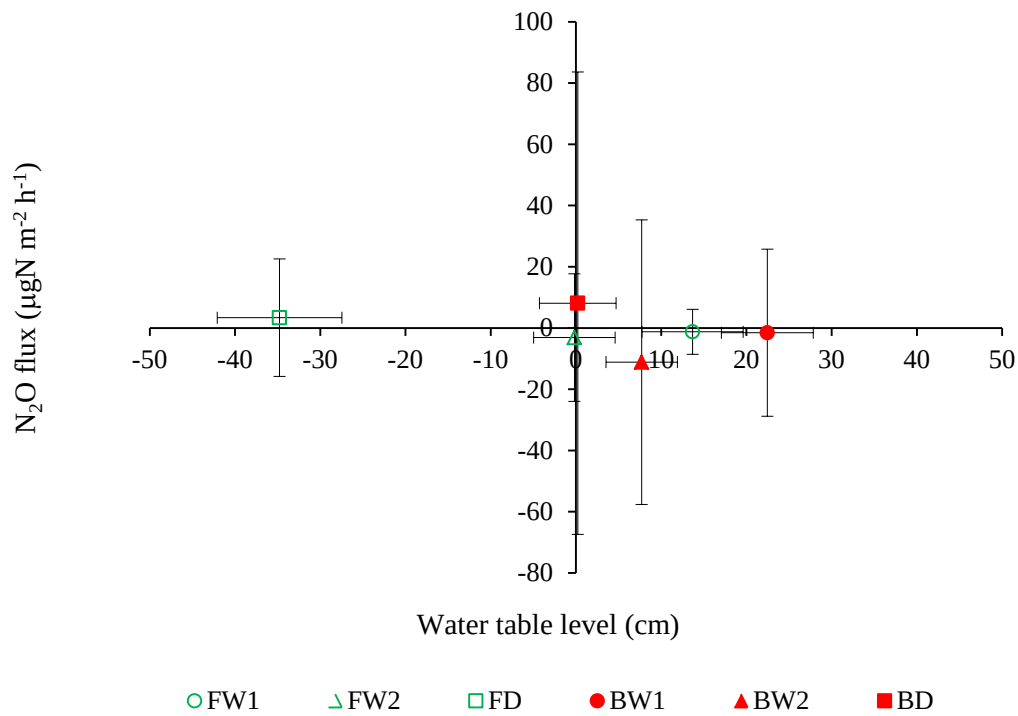


Fig. 5.7 Relationship between N₂O flux and water table level averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

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
Main factors and interaction						
Land use	1	7.34 **	1	6.33 *	1	4.13 *
Drainage	1	16.49 **	1	0.36	1	0.00
Land use*Drainage	1	6.05 *	1	6.64 *	1	8.59 **
Error	68		67		67	
Simple main effects						
Land use under flooded condition		15.85 **		0.00		0.47
Land use under drained condition		0.03		11.30 **		10.73 **
Drainage in forest vegetation		1.36		5.32 *		4.34 *
Drainage in burnt area		20.05 **		1.86		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**).

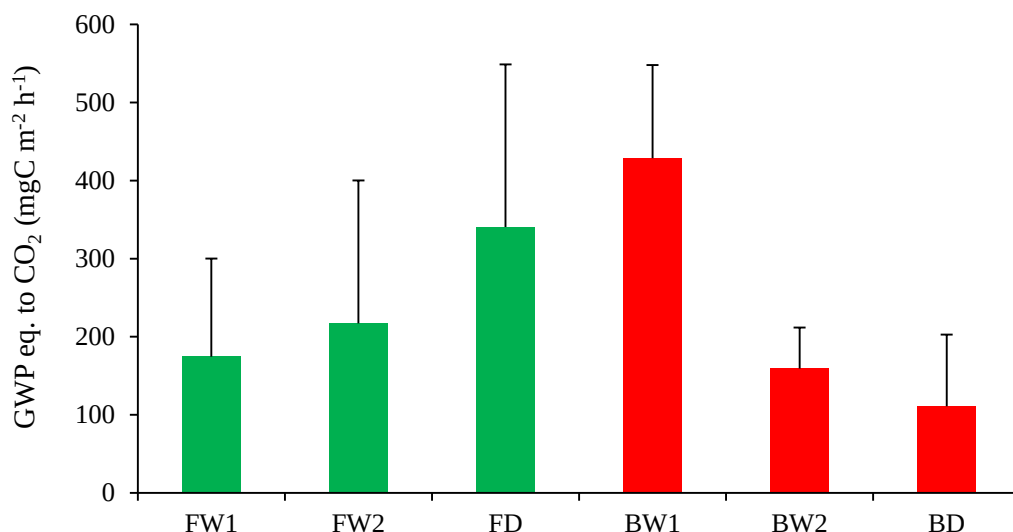


Fig. 5.8 GWP equivalent to CO₂-C flux at the ground surface averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

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 $\mu\text{gC m}^{-2} \text{h}^{-1}$ in a mixed peat swamp forest (water table level: -60 to -20 cm), from -7.44 to 102 $\mu\text{gC m}^{-2} \text{h}^{-1}$ in a sago plantation (-40 to 0 cm), and from -32.8 to 4.17 $\mu\text{gC m}^{-2} \text{h}^{-1}$ 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 ($5.75 \pm 6.66 \text{ mgC m}^{-2} \text{ h}^{-1}$) 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 ($0.0084 \pm 0.0321 \text{ mgC m}^{-2} \text{ h}^{-1}$), 2) CO₂ 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), and 3) N₂O fluxes ranged from –2.4 to –8.7 $\mu\text{gN m}^{-2} \text{ h}^{-1}$ in the flooded sites and from 3.4 to 8.1 $\mu\text{gN m}^{-2} \text{ h}^{-1}$ 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.

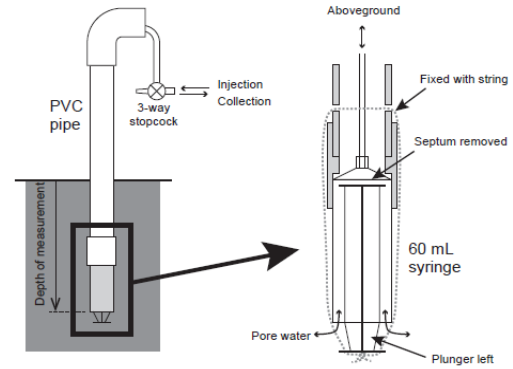


Fig. 6.3 Diagram of an in-situ equilibration chamber in the field.

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 $\pm 1.6^\circ\text{C}$ 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 ($124 \pm 84 \mu\text{mol L}^{-1}$) followed by the drained burnt site ($45.2 \pm 29.8 \mu\text{mol L}^{-1}$), the flooded forest sites ($1.15 \pm 1.38 \mu\text{mol L}^{-1}$), and the drained forest site ($0.860 \pm 0.819 \mu\text{mol L}^{-1}$) (**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.

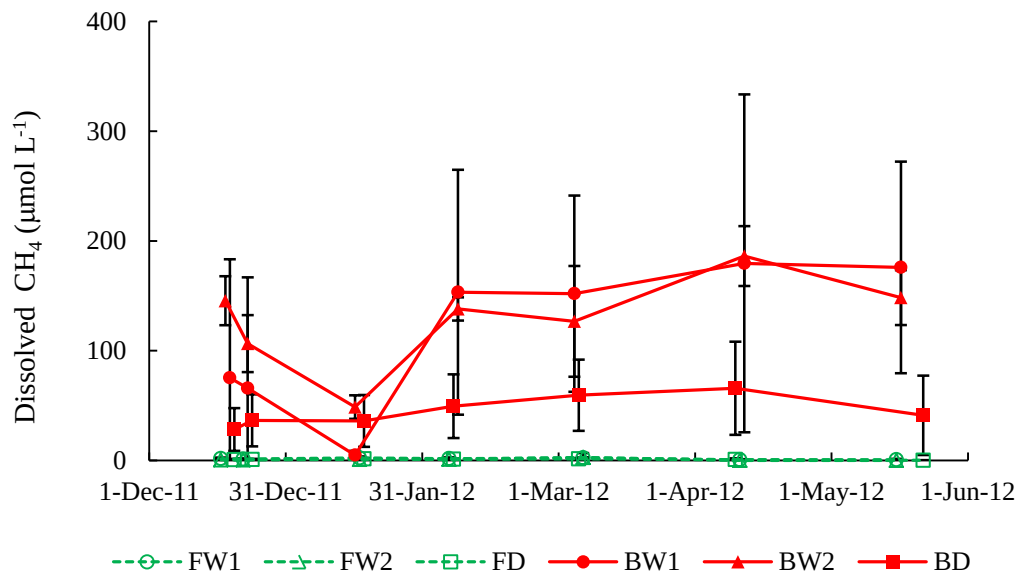


Fig. 6.4 Seasonal variations of dissolved CH₄ concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation.

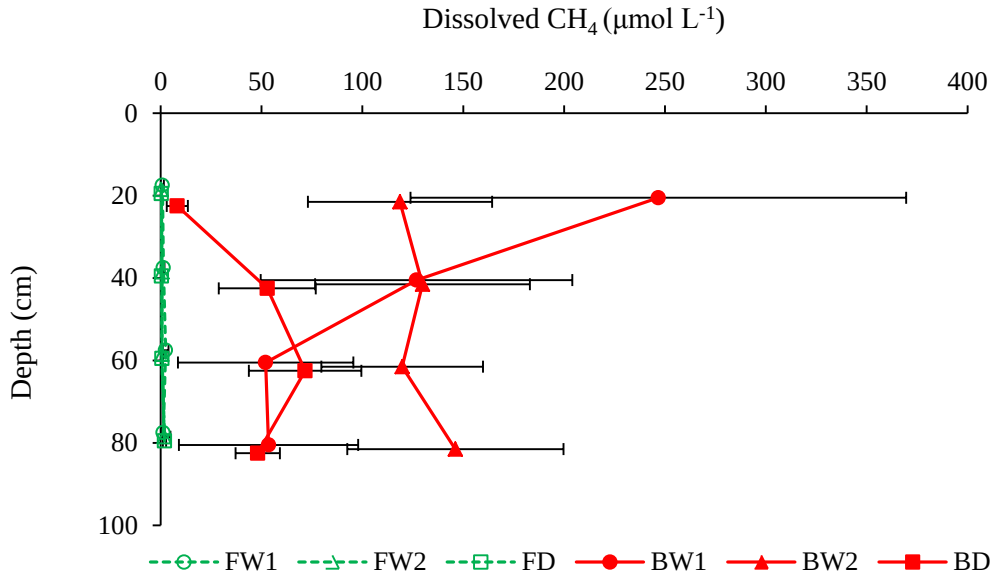


Fig. 6.5 Vertical distributions of dissolved CH₄ concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

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.

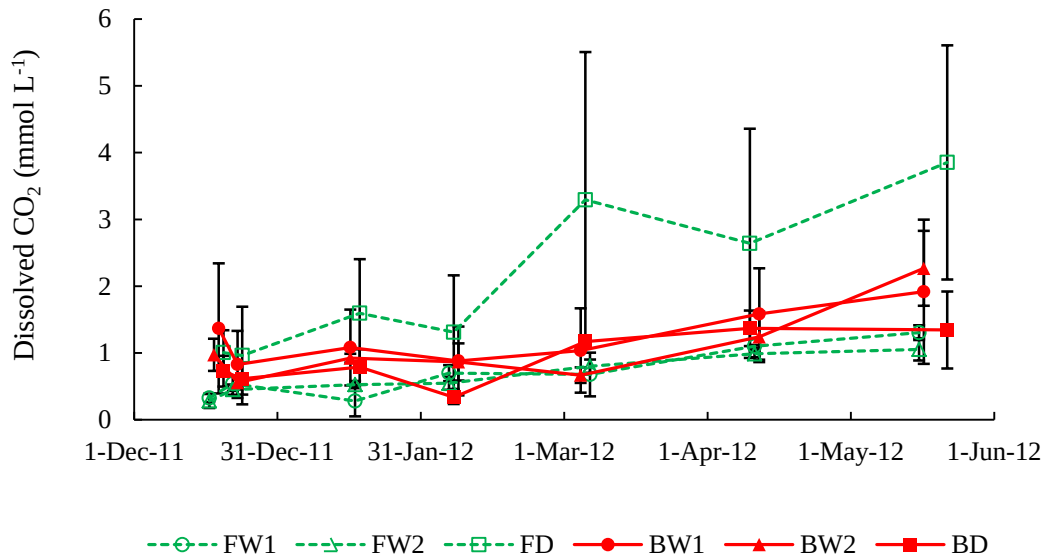


Fig. 6.6 Seasonal variations of dissolved CO₂ concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation.

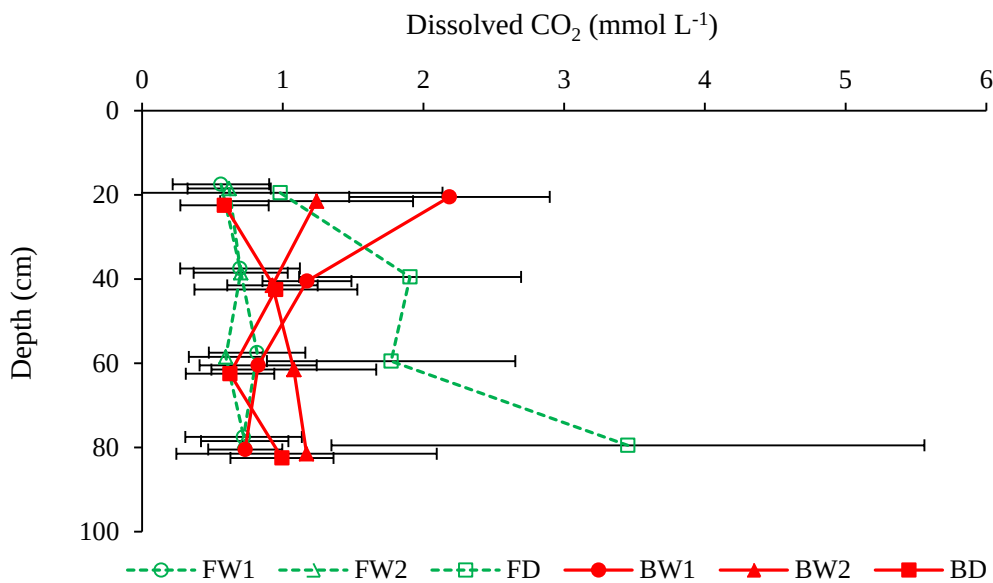


Fig. 6.7 Vertical distributions of dissolved CO₂ concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

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**).

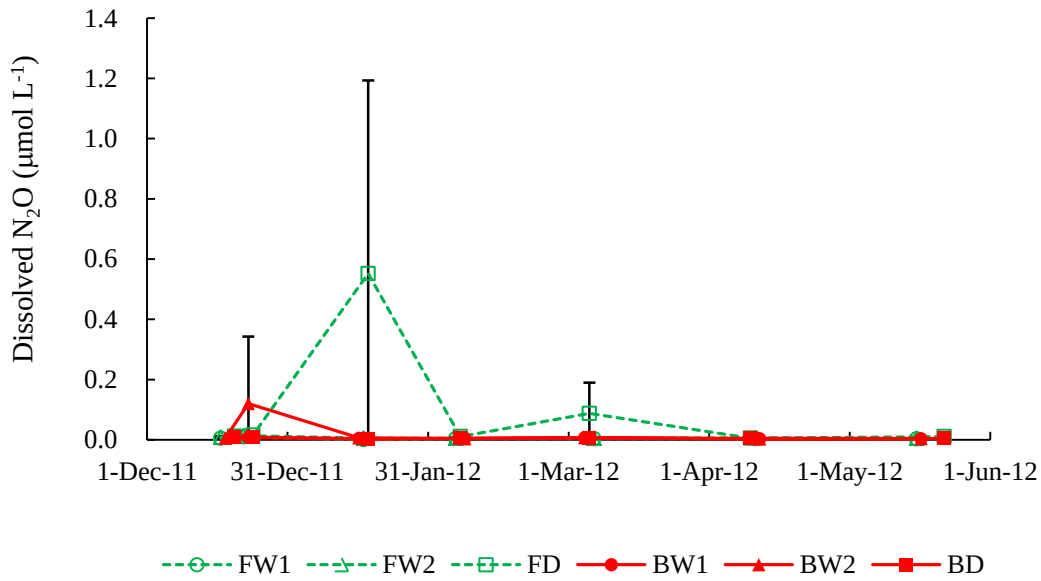


Fig. 6.8 Seasonal variations of dissolved N_2O concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation.

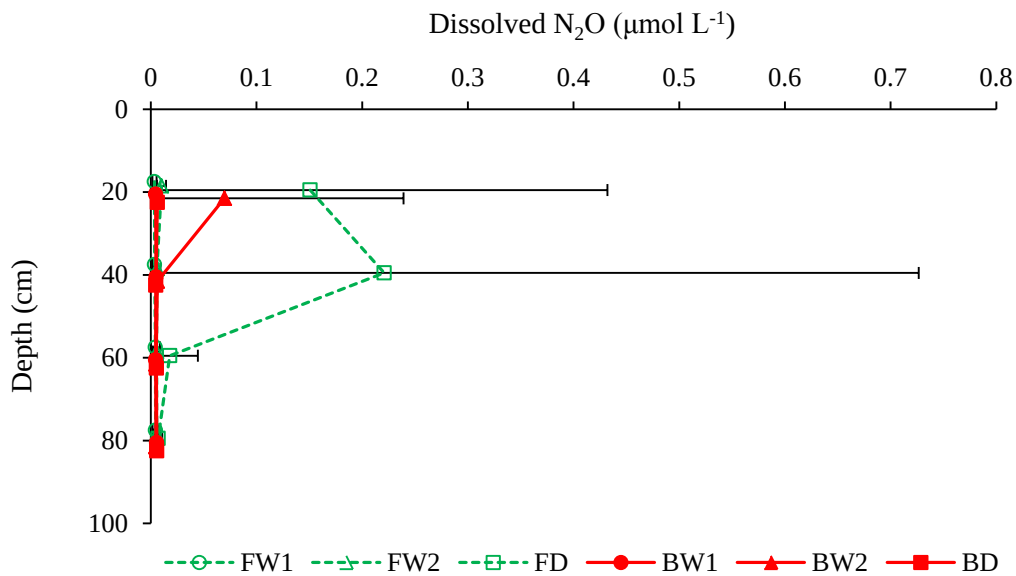


Fig. 6.9 Vertical distributions of dissolved N_2O concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.

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	
	<i>df</i>	<i>F</i>	<i>df</i>	<i>F</i>	<i>df</i>	<i>F</i>
<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 $\mu\text{mol L}^{-1}$, with average values of 48 and 226 $\mu\text{mol L}^{-1}$ 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 $\mu\text{mol L}^{-1}$. Terazawa *et al.* (2007) found that dissolved CH₄ concentrations in groundwater ranged from 5.6 to 28.4 $\mu\text{mol L}^{-1}$ 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 $\mu\text{mol L}^{-1}$, 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 ($124 \pm 84 \mu\text{mol L}^{-1}$) followed by the drained burnt site ($45.2 \pm 29.8 \mu\text{mol L}^{-1}$), the flooded forest sites ($1.15 \pm 1.38 \mu\text{mol L}^{-1}$), and the drained forest site ($0.860 \pm 0.819 \mu\text{mol L}^{-1}$). 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 ($6.9 \pm 5.6 \mu\text{mol L}^{-1}$) compared to the burnt sites ($4.0 \pm 2.9 \mu\text{mol L}^{-1}$) (**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 release 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; Ewelv *et al.*, 1987).

As with all forest, forested wetlands (swamps) are long-term sinks of atmospheric CO₂ in the absence of significant disturbances (Bridgham *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 have 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; Kozlowski, 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 Middelberg, 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 it's 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 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.

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