Effect of land use change and drainage on peat decomposition and greenhouse gas emission in a tropical peatland

Adji, Fengky Florante

2014-03-25

10.14943/doctoral.k11395

http://hdl.handle.net/2115/56102

theses (doctoral)

Fengky_Florante_Adji.pdf

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

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

FENGKY FLORANTE ADJI
ABSTRACT

To evaluate the hypotheses that 1) plant-mediated oxygen supply should oxidize methane (CH₄) produced in saturated tropical peat soil, and 2) this CH₄ oxidation should reduce total global warming potential (GWP) in a tropical peatland, the author compared the fluxes and dissolved concentrations of greenhouse gases [GHGs; CH₄, carbon dioxide (CO₂), and nitrous oxide (N₂O)] 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.

CH₄ fluxes were significantly influenced by land use and drainage, which were highest in the flooded burnt sites (5.75±6.66 mgC m⁻² h⁻¹) followed by the flooded forest sites (1.37±2.03 mgC m⁻² h⁻¹), the drained burnt site (0.22±0.143 mgC m⁻² h⁻¹), and the drained forest site (0.008±0.0321 mgC m⁻² h⁻¹). Dissolved CH₄ concentrations were also 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 (0.86±0.819 μmol L⁻¹). DO concentrations were influenced by land use only, which were significantly higher in the forest sites (6.9±5.6 μmol L⁻¹) compared to the burnt sites (4.0±2.9 μmol L⁻¹). These results suggested that CH₄ produced in the peat might be oxidized by plant-mediated oxygen supply in the forest sites. CO₂ fluxes were significantly higher in the drained forest site (340±250 mgC m⁻² h⁻¹ with the water table level of −20 to −60 cm) than in the drained burnt site (108±115 mgC m⁻² h⁻¹ with the water table level of −15 to +10 cm). Dissolved CO₂ concentrations were 0.6 to 3.5 mmol L⁻¹, also highest in the drained forest site. These results suggested the enhanced CO₂ emission by aerobic peat decomposition and plant respiration in the drained forest site. N₂O fluxes ranged from −2.4 to −8.7 μgN m⁻² h⁻¹ in the flooded sites and from 3.4 to 8.1 μgN m⁻² h⁻¹ in the drained sites. The negative N₂O fluxes might be caused by N₂O consumption by denitrification under flooded conditions. Dissolved N₂O concentrations were 0.005 to 0.22 μmol L⁻¹ but occurred at <0.01 μmol L⁻¹ in most cases.

GWP was mainly determined by CO₂ flux, with highest levels in the drained forest site.

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 and which was contributed on CH₄ oxidation under water saturated condition. The CH₄ flux and the dissolved CH₄ concentration at a depth of 20 cm in the drained burn 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 burn 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, and N₂O 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 CH₄ flux ......................................................................................................... 13
      2.3.1 Magnitude and variations of soil CH₄ flux ................................. 13
      2.3.2 Soils as a source or a sink of atmospheric CH₄ ...................... 14
      2.3.3 Environmental factors controlling CH₄ 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 CO₂ flux ......................................................................................................... 19
      2.4.1 Magnitude and variations of soil CO₂ flux ................................. 20
      2.4.2 Soil CO₂ flux and the ecosystems’ C balance ......................... 21
      2.4.3 Environmental factors controlling soil CO₂ flux ................... 22
      2.4.4 Autotrophic and heterotrophic respirations in soils ........... 22
  2.5 Soil N₂O flux .......................................................................................................... 23
      2.5.1 Magnitude of soil N₂O flux under different climate ................. 24
      2.5.2 Temporal and spatial variation in soil N₂O flux ....................... 25
      2.5.3 Environmental factors controlling N₂O 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 N₂O 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 CH₄ flux ................................................................. 64
    5.3.2 Soil CO₂ flux ................................................................. 66
    5.3.3 Soil N₂O flux ................................................................. 68
    5.3.4 GWP ............................................................................... 72
  5.4 Discussion ...................................................................................... 73
    5.4.1 Dynamics of soil CH₄ 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 CO₂ 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 N₂O 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 CH4 and N2O .................... 80
  6.2.2 Concentration of dissolved CO2 ................................. 81
  6.2.3 Calculation of dissolved GHG concentrations ......... 82
  6.2.4 Statistical analysis ......................................................... 83
6.3 Results .................................................................................. 84
  6.3.1 Dissolved CH4 concentration ................................. 84
  6.3.2 Dissolved CO2 concentration ................................. 86
  6.3.3 Dissolved N2O concentration .................................... 88
6.4 Discussion ........................................................................... 91
  6.4.1 Concentrations of dissolved CH4 in tropical peatlands . 91
  6.4.2 Concentrations of dissolved CO2 in tropical peatlands . 92
  6.4.3 Concentrations of dissolved N2O 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
<table>
<thead>
<tr>
<th>No.</th>
<th>Table caption</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1</td>
<td>Comparison of the fluxes and dissolved concentrations of CO₂, CH₄, and N₂O in wetland ecosystems in the world</td>
<td>39</td>
</tr>
<tr>
<td>4.1</td>
<td>Air temperature at all study sites averaged for the observation period (Dec. 2011 to May 2012)</td>
<td>53</td>
</tr>
<tr>
<td>4.2</td>
<td>DO concentration under different land use and drainage conditions averaged for the observation period</td>
<td>56</td>
</tr>
<tr>
<td>4.3</td>
<td>Results of the three-way ANOVA on DO concentrations among different land use, drainage, and depth</td>
<td>56</td>
</tr>
<tr>
<td>4.4</td>
<td>pH and EC at all study sites averaged for the observation period (Dec. 2011 to May 2012)</td>
<td>58</td>
</tr>
<tr>
<td>4.5</td>
<td>NO₃⁻ and NH₄⁺ concentrations at all study sites averaged for the observation period (Dec. 2011 to May 2012)</td>
<td>60</td>
</tr>
<tr>
<td>5.1</td>
<td>CH₄ flux measured at the ground surface under different land use and drainage conditions averaged for the observation period</td>
<td>66</td>
</tr>
<tr>
<td>5.2</td>
<td>CO₂ flux measured at the ground surface under different land use and drainage conditions averaged for the observation period</td>
<td>68</td>
</tr>
<tr>
<td>5.3</td>
<td>N₂O flux measured at the ground surface under different land use and drainage conditions averaged for the observation period</td>
<td>70</td>
</tr>
<tr>
<td>5.4</td>
<td>Summary of ANOVA on GHG fluxes and GWP under different land use and drainage conditions and multiple comparison among the study sites</td>
<td>71</td>
</tr>
<tr>
<td>5.5</td>
<td>GWP eq. to CO₂-C fluxes measured at the ground surface under different land use and drainage conditions averaged for the observation period</td>
<td>72</td>
</tr>
<tr>
<td>6.1</td>
<td>Dissolved CH₄ concentrations under different land use and drainage conditions averaged for the observation period</td>
<td>85</td>
</tr>
<tr>
<td>6.2</td>
<td>Dissolved CO₂ concentrations under different land use and drainage conditions averaged for the observation period</td>
<td>88</td>
</tr>
<tr>
<td>6.3</td>
<td>Dissolved N₂O concentrations under different land use and drainage conditions averaged for the observation period</td>
<td>90</td>
</tr>
<tr>
<td>6.4</td>
<td>Results of three-way ANOVA on the dissolved GHG concentrations among different land use, drainage, and depth</td>
<td>90</td>
</tr>
</tbody>
</table>
### LIST OF FIGURES

<table>
<thead>
<tr>
<th>No.</th>
<th>Figure caption</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1</td>
<td>Change in abundance of major GHGs for the last thousand years (IPCC, 2007)</td>
<td>4</td>
</tr>
<tr>
<td>2.2</td>
<td>Map of peatland distribution in the Southeast Asia (Page et al., 2004)</td>
<td>6</td>
</tr>
<tr>
<td>2.3</td>
<td>Overview of peat swamp forest canopy from the top of a tower</td>
<td>8</td>
</tr>
<tr>
<td></td>
<td>established in Sebangau area, Central Kalimantan - Indonesia (Image: Adji F.F)</td>
<td></td>
</tr>
<tr>
<td>2.4</td>
<td>Schematic representation of C cycle processes, flow paths, and stores</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td>in tropical peatland (after Jauhiainen et al., 2001)</td>
<td></td>
</tr>
<tr>
<td>2.5</td>
<td>Schematic illustration of drainage effects on a peatland dome</td>
<td>12</td>
</tr>
<tr>
<td></td>
<td>(modified from Delft Hydraulics, 2006)</td>
<td></td>
</tr>
<tr>
<td>2.6</td>
<td>Schemes of C fluxes and factors controlling them in the ecosystems</td>
<td>21</td>
</tr>
<tr>
<td>2.7</td>
<td>Diagram of the “hole-in-the-pipe” conceptual model describing soil</td>
<td>33</td>
</tr>
<tr>
<td></td>
<td>emissions of N₂O and NO (Davidson et al., 2000)</td>
<td></td>
</tr>
<tr>
<td>3.1</td>
<td>Locations of field observation sites</td>
<td>45</td>
</tr>
<tr>
<td>4.1</td>
<td>View of the tipping bucket rain gauge settled near the base camp of</td>
<td>48</td>
</tr>
<tr>
<td></td>
<td>Kalampangan canal</td>
<td></td>
</tr>
<tr>
<td>4.2</td>
<td>View of DO measurement at the study site</td>
<td>50</td>
</tr>
<tr>
<td>4.3</td>
<td>Diagram of an in-situ DO measurement</td>
<td>50</td>
</tr>
<tr>
<td>4.4</td>
<td>View of a set of water samplers for pH, EC, NO₃⁻, NH₄⁺, and DO</td>
<td>51</td>
</tr>
<tr>
<td></td>
<td>measurements at the study site</td>
<td></td>
</tr>
<tr>
<td>4.5</td>
<td>Seasonal changes in precipitation (bar) and water table level at each site</td>
<td>52</td>
</tr>
<tr>
<td></td>
<td>(line with symbol)</td>
<td></td>
</tr>
<tr>
<td>4.6</td>
<td>Seasonal variations in air temperature at each study sites</td>
<td>53</td>
</tr>
<tr>
<td>4.7</td>
<td>Vertical distributions in soil temperature at all soil depths. Error bars</td>
<td>54</td>
</tr>
<tr>
<td></td>
<td>show standard deviation. For better identifiability, the depths in the profiles</td>
<td></td>
</tr>
<tr>
<td></td>
<td>were slightly shifted from their actual depths (5, 10, 30, and 50 cm)</td>
<td></td>
</tr>
<tr>
<td>4.8</td>
<td>Vertical distributions of DO concentration in subsurface pore water</td>
<td>55</td>
</tr>
<tr>
<td></td>
<td>averaged for the observation period (Dec. 2011 to May 2012). Error bars show</td>
<td></td>
</tr>
<tr>
<td></td>
<td>standard deviation. For better identifiability, the depths in the profiles</td>
<td></td>
</tr>
<tr>
<td></td>
<td>were slightly shifted from their actual depths (10, 20, 40, and 80 cm)</td>
<td></td>
</tr>
</tbody>
</table>
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 NO₃⁻ 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 NH₄⁺ 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 CH₄ fluxes measure at the ground surface in the study sites. Error bars show standard deviation ................................. 65

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 ......................................................................................... 65

5.4 Seasonal variations of CO₂ fluxes measure at the ground surface in the study sites. Error bars show standard deviation ........................................ 67

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 ......................................................................................... 67

5.6 Seasonal variations of N₂O fluxes measure at the ground surface in the study sites. Error bars show standard deviation ................................. 69

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 ......................................................................................... 69

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 ......................................................................................... 72

6.1 View of the rapid equilibration method for the measurement of dissolved CH₄ and N₂O concentrations ....................................................... 81

6.2 View of field settings for dissolved CO₂ concentrations measurement at the study site ..................................................................................... 82

6.3 Diagram of an in-situ equilibration chamber in the field ..................... 82
6.4 Seasonal variations of dissolved CH$_4$ concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation ................................................................. 85

6.5 Vertical distributions of dissolved CH$_4$ concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation ................................................................. 86

6.6 Seasonal variations of dissolved CO$_2$ concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation ................................................................. 87

6.7 Vertical distributions of dissolved CO$_2$ concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation ................................................................. 87

6.8 Seasonal variations of dissolved N$_2$O concentrations in pore water averaged for depths of 20–80 cm at each site. Error bars show standard deviation ................................................................. 89

6.9 Vertical distributions of dissolved N$_2$O concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation ................................................................. 89
Acknowledgements

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

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

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

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

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

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

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

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

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

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

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

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

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

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

Many thanks to Mr. Untung Darung, Mr. Haiki Mart Yupi, Mr. Hendrik Segah Patianom and family, Mr. Rony Teguh and family, Mr. Atfritedy Limin, Ms. Nina Yulianti, Ms. Febrina Natalia Bahat, Mrs. Yulinda Wulandari, Mrs. Mandagi Ixchel Feibie and family, and also PPI H/S friends in Sapporo for their kindness and support.
Finally, my special thanks to beloved parents: Florante K. Adji and Herliany D. Gentong, parents-in-law: Matheus Ngadijono and (RIP) Herwaty HS. Tumon, Fredyantoni F. Adji (brother), Agus Hertanto and Guntoro Purwo Handoko (brother-in-law), Rina Mariana, Susi Dwi Hartanti, and Imelda (sister-in-law), also Tama, Nata, Gita (nephews), Santy Marlina (wife), Danika Ziva Angela (daughter), and Darren Floriartha Adji (son), for their understanding, support, and love that make my study abroad go on well.
Chapter 1

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 CH$_4$ fluxes in tropical peatlands was one-fifth that observed in temperate and boreal peatlands. They attributed these low CH$_4$ 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 CH$_4$ oxidized before entering the atmosphere ranges from 0 to up to >90% of the potential CH$_4$ 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 CH$_4$ concentrations (Fritz et al., 2011; Liebner et al., 2012; van der Nat and Middelburg, 1998). Thus, if plant-mediated oxygen transport and CH$_4$ oxidation also occur in tropical peatlands, significant differences in soil CH$_4$ fluxes, dissolved CH$_4$ 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 CH$_4$ dynamics at the vegetated and non-vegetated lands in tropical peatland, the following hypotheses are addressed: 1) DO supplied from plant roots should oxidize CH$_4$ produced in saturated tropical peat soil, and 2) this CH$_4$ 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 CH$_4$, CO$_2$, and N$_2$O, 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 CH$_4$, CO$_2$, and N$_2$O 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 CH$_4$, CO$_2$, and N$_2$O 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.
Chapter 2

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±0.31°C and 0.52±0.13°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 (~30,000–24,000 $^{14}$C yrs BP). Accumulation was most rapid in the early Holocene (~9600–7000 $^{14}$C yrs BP, ~11,000–8000 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 with in 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 (~0.1 g cm$^{-3}$) compared to mineral soils, being formed of ~10% tree remains and 90% water (Hooijer et al., 2010), and is 50–60% 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 Andriesse, 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 in 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; Wösthen 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 growis nutrient poor (Andriesse, 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 (Andriesse, 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 itself; Fig. 2.4). In addition, smaller amount of C is leached out from the system in drainage runoff as dissolved organic C (DOC) or particulate organic C (POC; Moore et al., 2011).

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

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

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 (Bridgham et al., 2006; Clymo et al., 1998).
2.2.4 Development of tropical peatlands for agriculture

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

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

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

Deforestation and development of peatland usually accompanied the drainage of water from that areas, so much of the C has been lost by aerobic peat decomposition. In addition, the water level was likely to become uncontrollable during an exceptionally long dry period, which was frequently observed in El Niño and Southern Oscillation (ENSO) years (Hirano et al., 2007 and 2012; van der Werf et al., 2008), and to make the peat severely dry and susceptible to fire. Drained organic soils for agriculture have a special significance in the atmospheric N₂O load (Flessa et al., 1998; Kasimir-Klemedtsson et al., 1997; Maljanen et al., 2003b). As drainage increases peat mineralization rates and NO₃⁻ availability, the potential
for N\textsubscript{2}O production becomes significantly greater (Hadi et al., 2000; Jauhiainen et al., 2012). In addition, applications of N fertilizer generally serve to increase N\textsubscript{2}O 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 O\textsubscript{2} 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 CH\textsubscript{4} and CO\textsubscript{2} 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 CO\textsubscript{2} 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 $-600$ gC m$^{-2}$ yr$^{-1}$ ($-6$ tC ha$^{-1}$ 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;
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 $4.02 \times 10^6$ J m$^{-3}$ K$^{-1}$ versus that of dry peat, which is approximately $0.58 \times 10^6$ J m$^{-3}$ K$^{-1}$ (Oke, 1987). Therefore, drier conditions could indirectly stimulate decomposition rates by affecting thermal regimes (Laine et al., 2006; Lieffers, 1988). Increased nutrient availability with faster soil mineralization rates under warmer, drier soil conditions also could have consequences for vegetation.

2.3 Soil CH$_4$ flux

Atmospheric concentrations of CH$_4$ 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 CH$_4$ production arises from anthropogenic sources and about 30% from natural sources. Currently the estimations of CH$_4$ potential from tropical peatland/swamps, upland soils, oceans, and lakes are 127.6, –30, 9.1, and 30 TgCH$_4$ yr$^{-1}$, respectively (EPA, 2010). The actual amount of CH$_4$ emitted to the atmosphere depends on the balance between CH$_4$ production and consumption as well as the CH$_4$ transport efficiency (Couwenberg, 2009). CH$_4$ is removed from the atmosphere by reacting with the hydroxyl radical (OH) and is ultimately converted to CO$_2$. Minor removal processes also include reaction with Cl$^-$ in the marine boundary layer, a soil sink, and stratospheric reactions. Increasing emissions of CH$_4$ reduce the concentration of OH, a feedback that may increase methane’s atmospheric lifetime (IPCC, 2001).

2.3.1 Magnitude and variations of soil CH$_4$ flux

Soils which are permanently flooded are strong sources for atmospheric CH$_4$ in all climate zones of the world. The magnitude of fluxes from such ecosystems is in the range of <5 to <200 mgCH$_4$ m$^{-2}$ d$^{-1}$ and can exceed 20 mgCH$_4$ m$^{-2}$ h$^{-1}$ (Delmas et al., 1992; Moore et al., 1990a and b; Scott et al., 1999). CH$_4$ 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 CH$_4$
uptake rates by soils are mostly <300 μgCH₄ m⁻² h⁻¹ and are approximately one magnitude lower than the rates of CH₄ 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 CH₄ fluxes between soils and the atmosphere. As at the moment, the IPCC estimate of CH₄ 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 CH₄ sink (−0.208 to −0.368 g m⁻² yr⁻¹; −13.0 to −22.9 mmol CH₄ m⁻² yr⁻¹) and the deforested site a comparable CH₄ source (0.197 to 0.275 g m⁻² yr⁻¹; 12.3 to 17.1 mmol CH₄ m⁻² yr⁻¹) (Jauhiainen et al., 2008).

CH₄ 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 CH₄ 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 CH₄ emissions. In recent years, biotic factors such as vegetation are considered to control CH₄ emissions from wetlands, because aquatic plants affect the production, consumption, and transport of CH₄ (Joabsson and Christensen, 2001; Whitting and Chanton, 1992). For example, in the Luanhaizi wetland, spatial variation of CH₄ 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 CH₄ emissions generally being higher from wetter soils, and little temporal variations in CH₄ 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 CH₄ fluxes are influenced by latitude, with artic, boreal, and some temperate regions characterized by pronounced CH₄ 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 CH₄

Soils can function as net sinks or as net sources for atmospheric CH₄. The exchange of CH₄ between soils and the atmosphere is the net result of simultaneous production of CH₄ in
predominantly anaerobic zones of the soil and the oxidation of CH$_4$ in predominantly oxic zones of the soil. Production and consumption of CH$_4$ within the soil profile is controlled by several environmental factors: O$_2$ and substrate availability, soil properties, and climate. The primary control if soils are sinks or sources for the atmospheric CH$_4$ is O$_2$ availability in the soil profile.

In waterlogged soils, CH$_4$ 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 CO$_2$-H$_2$ (Knowles, 1993). The availability of these substrates and the magnitude of CH$_4$ production are suggested by the net primary production in an ecosystem (Aselmann and Crutzen, 1989). Further controllers of CH$_4$ production are soil pH (Knowles, 1993) and temperature (Conrad, 1996).

Since CH$_4$ 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 CH$_4$ can be oxidized by methanotrophic bacteria. Several studies in wetland ecosystems have shown that 40 to 95% of total CH$_4$ 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 CH$_4$ 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 CH$_4$ 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 CH$_4$ (Adamsen and King, 1993; Castro et al., 1995; Crill, 1991; Keller et al., 1983; Yavitt et al., 1990). Main uptake activity of atmospheric CH$_4$ 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 CH$_4$ oxidation activity or may even show low rates of net-production of CH$_4$ (Adamsen and King, 1993; Sexstone and Mains, 1990). The reasons for this wide spread observations that the main CH$_4$ 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 CH$_4$-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 NH$_4^+$ content in the organic layer as compared to the uppermost mineral layer inhibit CH$_4$ 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 CH$_4$ activity may be due to a competitive inhibition of the enzyme CH$_4$ monooxygenase by NH$_3^+$ (Adamsen and King, 1993; Dunfield and Knowles, 1995; King and Schnell, 1994) or the production of toxic products (NO$_2^-$ and NH$_2$OH) produced during enhanced NH$_4^+$ oxidation (King and Schnell, 1994). Furthermore, the organic layer may contain further compounds that may control CH$_4$ oxidation activity. As shown by Amaral and Knowles (1998), CH$_4$ oxidation rates can be inhibited by increasing concentrations of monoterpene. They had suggested that soil monoterpene distribution may be responsible for stratification of CH$_4$ oxidation activity in soils.

2.3.3 Environmental factors controlling CH$_4$ flux

Empirical models have been tried to explain the variation in CH$_4$ 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 CH$_4$ 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; Spheppard 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 CH$_4$ following in situ investigation.

2.3.3.1 Land use

CH$_4$ 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; Gulledge et al., 1997; Sitaula et al., 1995; Steudler et al., 1989; Whalen and Reeburg, 2000) whereby the rates of atmospheric CH$_4$ oxidation is strongly reduced by increasing N availability. CH$_4$ 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 CH$_4$ 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 CH$_4$ (Keller and Reiner, 1994). In contrast to the short-term effects of land use change from forestry to agriculture on CH$_4$ fluxes, it may take decades after conversion of arable land to forest to regain CH$_4$ 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 CH$_4$ emission rates by serving as a conduit for gas by means of aerenchyma system by which CH$_4$ produced in anaerobic environments bypasses the oxidation zone where less than 10% of CH$_4$ 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 CH$_4$ emission (Banker et al., 1995; Holzapfel-Pschorr et al., 1986; Jia et al., 2001; Yagi et al., 1996). Besides transporting CH$_4$, plants can also stimulate CH$_4$ production by providing substrates for methanogens and facilitate CH$_4$ oxidation by delivering O$_2$ downward into the rhizosphere for methanotrophs (Jia et al., 2001; van der Nat and Midderberg, 1998). Moreover, plant species also greatly affected the magnitude of CH$_4$ 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 CH$_4$ emissions than direct manipulation of air and soil temperature.

Up to date, the majority of tree species that possess adaptive structures to facilitate O$_2$ ingress are also capable of mediating CH$_4$ egress. The previous studies have shown six of species tree: Elaocarpaceae (Elaeocarpus mastersii), Ebenaceae (Diospyros bantamensis), Myrtaceae (Tristaniopsis sp. 2), Clusiaceae or Guttiferae (Mesua sp. 1), Lauraceae (Litsea elliptica), Annonaceae (Xylopia fusca) emitted CH$_4$ 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 CH$_4$ dissolved in pore water explain up to 80% of stem CH$_4$ flux variations.
2.3.3.2 Drainage

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

Lowering of water table in peatlands has occurred over a large area in temperate, boreal and subarctic regions (Armentano and Menges, 1986; Gorham, 1991). It would be expected that this would increase the thickness of the zone of CH4 consumption and decrease the thickness of the zone of CH4 production in the peat profile, resulting in a reduction in CH4 emission rates. Thus drained peatlands have very small CH4 consumption rates (generally <2 mg CH4 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 CH\textsubscript{4} oxidation will increase with increasing temperatures. Several studies have shown that soils under $<10^\circ\text{C}$ are subject to temperature change (Castro \textit{et al.}, 1995; Crill, 1991; Steinkamp \textit{et al.}, 2001). At temperatures $>10^\circ\text{C}$, other factors such as soil moisture have been observed to be more important for the magnitude of CH\textsubscript{4} uptake than soil temperature (Castro \textit{et al.}, 1995; Steinkamp \textit{et al.}, 2001).

Atmospheric CH\textsubscript{4} uptake is largely controlled by gas diffusion resistance within the soil. Therefore, the structure of the organic layer (Brumme and Borken, 1999; Dong \textit{et al.}, 1998) and the texture of the mineral soil (Boeckx \textit{et al.}, 1997; Dorr \textit{et al.}, 1993) have a huge impact on the magnitude of rates of CH\textsubscript{4} oxidation in soils.

Moisture availability appears to act as a switch turning CH\textsubscript{4} emissions on and off. The presence of a large saturated zone creates a high potential for methanogenes. 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 CH\textsubscript{4} oxidation activity by limiting substrate availability (Brumme and Borken, 1999; Castro \textit{et al.}, 1995; Gulledge and Schimel, 1998). Furthermore, high soil moisture contents will also decrease O\textsubscript{2} availability in the soil profile, thereby inhibiting the process of CH\textsubscript{4} oxidation, since for the oxidation of CH\textsubscript{4} to methanol by the enzyme CH\textsubscript{4} monoxygenase, O\textsubscript{2} is a prerequisite (Gulledge and Schimel, 1998; Knowles, 1993).

2.4 Soil CO\textsubscript{2} flux

The atmospheric CO\textsubscript{2} 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 \textit{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 CO\textsubscript{2} increase is caused by anthropogenic emissions of CO\textsubscript{2} (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 \textit{et al.}, 2000; Pete Smith, 2004; Valentini \textit{et al.}, 2000). In its second assessment, IPCC (1996) also stated that the increased amount of CO\textsubscript{2} 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 CO$_2$ flux

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

Soil CO$_2$ 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 CO$_2$ 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, CO$_2$ fluxes at the forest soil were clearly higher in comparison to the deforested area. Cumulative forest floor CO$_2$ fluxes (7305 to 7444 g m$^{-2}$ yr$^{-1}$; 166.0 to 1.69.2 molCO$_2$ m$^{-2}$ yr$^{-1}$) and deforested site CO$_2$ fluxes (2781 to 2608 g m$^{-2}$ yr$^{-1}$; 63.2 to 59.3 molCO$_2$ 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 CO$_2$ fluxes from tropical peatlands were 109.0±9.0 mol m$^{-2}$
yr\(^{-1}\) (4794±422 gCO\(_2\) m\(^{-2}\) yr\(^{-1}\)), and the soil CO\(_2\) flux increased exponentially with soil temperature (\(T_s\)) even within an-amplitude of 4–5°C.

### 2.4.2 Soil CO\(_2\) flux and the ecosystems’ C balance

Soil CO\(_2\) 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 CO\(_2\) to the atmosphere through respiration and consume CO\(_2\) from the atmosphere via photosynthesis. CO\(_2\) is assimilated in photosynthesis by trees and ground vegetation and translocation to soils through several pathways (Fig. 2.6). Significant amounts of C are allocated to the root systems for root growth and root maintenance. When the roots die, the C is added to the soil as dead organic matter. Carbon is also added to the ground surface and humus from aboveground biomass through litter fall and leaching of dissolved organic matter from the canopy (Edwards and Harris, 1977; Kalbitz et al., 2000) and from roots (Högberg et al., 2001). Carbon is released from the soil to the atmosphere through the decomposition of dead organic matter and through the respiration by roots, root mycorrhizal fungi and other soil micro-organisms (Chapin III and Ruess, 2001; Gaudinski et al., 2000). Some of the C is also leached out of the ecosystem dissolved in ground water especially in peatlands (Urban et al., 1989; Sallantaus, 1992).

![Fig. 2.6](https://example.com/fig26.png)  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 CO\(_2\). In a climax ecosystem, soil respiration is taken to be approximately balanced by photosynthetetic uptake of CO\(_2\) 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 CO$_2$ 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 CO$_2$ flux

The conversion of forests to agriculture is responsible for a substantial increase in atmospheric CO$_2$ 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 CO$_2$ 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 CO$_2$ emissions from peatlands, because saturated soils limit the diffusion of atmospheric O$_2$ into the peat, limiting microbial activity and decomposition rates. Conversely, a water table decline increases O$_2$ diffusion into soils allowing aerobic decomposition, which increases CO$_2$ 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 CO$_2$ efflux due to sensitivity of soil respiration in elevated temperatures (Boone et al., 1998). While liming activities increased the CO$_2$ emission to 4.1 tC ha$^{-1}$ yr$^{-1}$ from a temperate forest and drastically reduced the N$_2$O emission to 1.5 kgN ha$^{-1}$ yr$^{-1}$ (Brumme and Beese, 1992).

2.4.4 Autotrophic and heterotrophic respirations in soils

Soil CO$_2$ emission, as the result of soil respiration generates mainly from autotrophic (root) and heterotrophic (microbial) activity (Janssens et al., 2001). Soil microorganism release CO$_2$ 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 Kingerlee, 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 N$_2$O 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 N$_2$O, 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 N$_2$O flux is the result of dynamic production and consumption processes in soil. However, most of the N$_2$O produced within the soil column may be consumed and never reach the atmosphere (Arah et al., 1991; Seiler and Conrad, 1981). Consumption of N$_2$O is due to the reduction to N$_2$. Oxidation of N$_2$O 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 N$_2$O 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 N$_2$O 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 N$_2$O from nitrification and denitrification.

2.5.1 Magnitude of soil N$_2$O flux under different climate

The total flux of N$_2$O 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 N$_2$O emissions come from natural sources, but individual source estimates remain subject to significant uncertainties (Forster et al., 2007). According to EPA (2010), N$_2$O 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 N$_2$O 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) TgN$_2$O 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 N$_2$O was from about 0.8 to about 2.1 TgN$_2$O-N, respectively (Mosier and Zhaoliang, 2000).
There is general agreement that tropical forest soils are very important \( \text{N}_2\text{O} \) 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 \( \text{N}_2\text{O} \) emission from organic soil is mostly done in boreal and temperate regions. According to the current IPCC Guidelines, \( \text{N}_2\text{O} \) emissions from organic soils in cool and temperate climates are estimated to be 8 kg\( \text{N}_2\text{O}-\text{N} \) ha\(^{-1}\). Mineralization rates of \( \text{N} \) 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 kg\( \text{N}_2\text{O}-\text{N} \) ha\(^{-1}\) (Penman et al., 2000).

The increasing use of inorganic \( \text{N} \) fertilizers in Indonesia having a warm, tropical climate with frequent rainfall events may contribute to \( \text{N}_2\text{O} \) production significantly. Previous studies have explained that the annual \( \text{N}_2\text{O} \) emissions from the croplands in tropical peatland were significantly higher (21±5.4 to 131±59 kg\( \text{N}_2\text{O}-\text{N} \) ha\(^{-1}\) yr\(^{-1}\) in 2002–2003; 52±8.2 to 259±44 kg\( \text{N}_2\text{O}-\text{N} \) ha\(^{-1}\) yr\(^{-1}\) in 2003–2004) compared to the natural forest (0.62±0.11 kg\( \text{N}_2\text{O}-\text{N} \) ha\(^{-1}\) yr\(^{-1}\) in 2002–2003; 4.4±1.2 kg\( \text{N}_2\text{O}-\text{N} \) ha\(^{-1}\) yr\(^{-1}\) in 2003–2004), regenerated forest (0.40±0.32 kg\( \text{N}_2\text{O}-\text{N} \) ha\(^{-1}\) yr\(^{-1}\) in 2002–2003; 4.0±1.9 kg\( \text{N}_2\text{O}-\text{N} \) ha\(^{-1}\) yr\(^{-1}\) in 2003–2004), and burned forest (0.97±0.65 kg\( \text{N}_2\text{O}-\text{N} \) ha\(^{-1}\) yr\(^{-1}\) in 2002–2003; 1.5±0.7 kg\( \text{N}_2\text{O}-\text{N} \) ha\(^{-1}\) yr\(^{-1}\) in 2003–2004) (Takakai et al., 2006). In addition, the \( \text{N}_2\text{O} \) emissions in conventionally cultivated croplands and unplanted/unfertilized treatments (bare soil) in each cropland at the tropical peatland varied from 10.9 to 698 kg\( \text{N} \) ha\(^{-1}\) yr\(^{-1}\) and 6.55 to 858 kg\( \text{N} \) ha\(^{-1}\) yr\(^{-1}\), respectively (Toma et al., 2011).

### 2.5.2 Temporal and spatial variation in soil \( \text{N}_2\text{O} \) flux

Most studies in temperate soils had reported daily variations of \( \text{N}_2\text{O} \) 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 \( \text{N}_2\text{O} \) (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 N$_2$O emissions at all. The reason for these conflicting results remains to be clarified.

With respect to tropical forest ecosystems it was generally found that N$_2$O 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 N$_2$O 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 N$_2$O emissions (Dobbie et al., 1999; Flessa et al., 1998).

N$_2$O 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 results 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 N$_2$O 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 N$_2$O 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, NO$_3^-$, and NH$_4^+$ (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 N$_2$O fluxes because it has a strong influence on the hydrological and pedological progresses 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 N$_2$O emission from tropical rain forest soils (Breuer et al., 2000). The spatial variability of N$_2$O 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 N$_2$O 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 N$_2$O budget.

2.5.3 Environmental factors controlling N$_2$O flux

There are several environmental factors that influence the N$_2$O emission from agroecosystems (Mosier et al., 1996). The processes that may be involved in determining the flux between soil and atmosphere of N$_2$O 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 N$_2$O (Bremner et al., 1980). Under most soil conditions, however, the turnover of N$_2$O is caused by biochemical reactions taking place within microbial cells. Thus it is obvious that microbial processes ultimately cause N$_2$O flux between the soil and the atmosphere.

Previous studies have been described about the regulation of N$_2$O turnover in soil. Generally, continuous labile N availability from vegetation or labile organic compounds, water table level, soil temperature, soil moisture, and soil O$_2$ concentrations are the universal controlling factors for N$_2$O 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 N$_2$O 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 N$_2$O. 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 O$_2$ thus enhancing anaerobic microbial metabolism and possibly stimulating N$_2$O production. On the other hand, however, diffusion of N$_2$O is also impeded and thus consumption of this gas is stimulated.

NO$_3^-$ 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 N$_2$O trapped in the soil (Machefert et al., 2004). Some researchers have reported enormous pulses of N$_2$O 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 N$_2$O (Davidson et al., 1991 and 1993; Serca et al., 1998).

Soil temperature influences the physiological activity of soil microorganism and it can alter microbial activity in soil, rates of organic decomposition and ecosystem C storage (Zak et al., 1999). The enhanced microbial respiration and N mineralization are related to larger substrate pools and the increases in soil temperature (MacDonald et al., 1995; Zogget al., 1999). Thus these activities will influence the nitrification and denitrification processes, and thereby biotic formation of N$_2$O (Keeney et al., 1979; Malhi et al., 1990). Soil temperature is also responsible for the rate of gas diffusion from soils and influences the N$_2$O/N$_2$ 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 N$_2$O 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 N$_2$O emission was independent from changes in soil temperature indicating that other environmental factors, e.g. soil moisture, were stronger modulators for N$_2$O than soil temperature.

Soil temperature could be the main determinants for N$_2$O 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 N$_2$O emission over a whole year in tropical peatland in South Kalimantan. N$_2$O 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 N$_2$O emission. They explained this result as inhibition of N$_2$O 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; Gödde 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 N\textsubscript{2}O 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 NH\textsubscript{4}\textsuperscript{+} and NO\textsubscript{3}\textsuperscript{−} might be a minor significance for the prediction of N trace gas fluxes (Skiba et al., 1998) and that actual microbial N turnover rates might be better predictors of in situ N\textsubscript{2}O flux (Gasche and Papen, 1999; Papen and Butterbach-Bahl, 1999). A conceptual model has been proposed in which various environmental regulators affect ammonium (NH\textsubscript{4}\textsuperscript{+}) oxidation and NO\textsubscript{3}\textsuperscript{−} reduction in a hierarchical way (Robertson et al., 1989; Tiedje, 1988).

2.5.3.2 Biotic factors and human-induced influences

The main N\textsubscript{2}O 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 N\textsubscript{2}O 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 N\textsubscript{2}O emission (Papen and Butterbach-Bhal, 1999). Also for tropical rain forest in Australia, the mineral soil was identified to represent the most important soil layer for N\textsubscript{2}O 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 N\textsubscript{2}O 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 N\textsubscript{2}O emission rates than sites lacking these N fixing tree species (Erickson et al., 2001).

The enhanced N\textsubscript{2}O emissions from natural and agricultural ecosystems are believed to be caused by increasing soil N availability driven by increased fertilizer application, agricultural N\textsubscript{2} 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 N\textsubscript{2}O. However, animal production systems, biological N fixation and erosion and leaching from soil systems are also the important contributors to N\textsubscript{2}O emission (Galloway et al., 2003; Skiba et al., 1998). Drainage alone increases the N\textsubscript{2}O emissions dramatically, but if the drained soil is used for agriculture, fertilization enhance the N\textsubscript{2}O emission even more (Kasimir-Klemedtsson et al., 1997). Fertilization and the form of N fertilizer affect the N\textsubscript{2}O 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 N\textsubscript{2}O emissions is still subject of study (Eichner, 1990; Stehfest and Bouwman, 2006), it has been observed that NO\textsubscript{3}\textsuperscript{-} 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 N\textsubscript{2}O emissions up to 25.7 times (Trujillo-Tapia et al., 2008).

2.5.4 Nitrification and denitrification for N\textsubscript{2}O dynamics

The N\textsubscript{2}O 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 N\textsubscript{2}O flux. Both denitrification and nitrification processes form and release N\textsubscript{2}O, 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).

N\textsubscript{2}O 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 N\textsubscript{2}O is an obligatory intermediate during denitrification. Denitrification is the sequential reduction of NO\textsubscript{3}\textsuperscript{-} \(\rightarrow\) NO\textsubscript{2}\textsuperscript{-} \(\rightarrow\) NO \(\rightarrow\) N\textsubscript{2}O \(\rightarrow\) N\textsubscript{2}, 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 O\textsubscript{2} (Tiedje, 1988). Some denitrifiers can denitrify even in the presence of O\textsubscript{2} (Robertson and Kuenen, 1988). If soils containing NO\textsubscript{3}\textsuperscript{-} become anaerobic, the availability of organic C that enhances the activity of denitrifiers is the limiting factor the reduction of NO\textsubscript{3}\textsuperscript{-}. 
Nitrification may be a significant source of N\textsubscript{2}O 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\textsubscript{2} as the principal C source for biomass formation. N\textsubscript{2}O production is enhanced in soils having a high mineralization capacity to form NH\textsubscript{4}\textsuperscript{+} or treated with nitrifiable forms of N.

In most forest soils, the importance of nitrification or denitrification as the main source of N\textsubscript{2}O 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\textsubscript{2} demand caused by the presence of easily mineralizable organic matter. Though denitrification has been demonstrated to be a very important source of N\textsubscript{2}O 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\textsubscript{2}O 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\textsubscript{2}O in soil seems to be produced during NH\textsubscript{4}\textsuperscript{+} oxidation and NO\textsubscript{3}\textsuperscript{−} reduction, the exact production mechanism is usually not easily understood for a specific field situation.

Release of N\textsubscript{2}O 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\textsubscript{2}O 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\textsubscript{2}O 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\textsubscript{4}\textsuperscript{+} oxidation by nitrifying bacteria and NO\textsubscript{3}\textsuperscript{−} reduction by denitrifying bacteria; and

2. The size of the holes in the pipe through which gaseous N\textsubscript{2}O, NO, and N\textsubscript{2} leak (or the amount of N that leaks out of the pipe) which is determine 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$_2$O and NO (Davidson et al., 2000).

The first level of regulation determines the total amount of N oxides produced (NO + N$_2$O) while the second level of regulation determines the relative importance of NO and N$_2$O 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$_2$O. GWP values for emissions of CO$_2$, CH$_4$, and N$_2$O on a molar basis for a 100-year time horizon are 1:25:298. Although CH$_4$ and especially N$_2$O 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 $\rightarrow$ concentration changes $\rightarrow$ radiative forcing $\rightarrow$ climate impacts $\rightarrow$ societal and ecosystem impacts $\rightarrow$ economic “damage” (O’Neill, 2000; Smith and Wigley, 2000; Fuglestvedt 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 CO$_2$ and CH$_4$ at depth in boreal and temperate peatlands. Previous studies have shown that under the field conditions CO$_2$ and CH$_4$ concentrations in the peat are mostly relatively uniform at levels of 1–10 mmol L$^{-1}$ for CO$_2$ and 0.1–1 mmol L$^{-1}$ for CH$_4$ (Nilsson and Bohlin, 1993). In addition, Blodau and Moore (2003) also reported CH$_4$ 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 CO$_2$ and CH$_4$ 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 CO$_2$ fluxes and dissolved CO$_2$ concentrations. The dissolved CO$_2$ 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 cm$^2$ 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 CO$_2$ 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 CO$_2$ 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, Sebacher 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$_2$O in soil solutions are also of very high temporal and spatial variability (Bowden and Bormann, 1986; Dowdell et al., 1979; Minami and Fukushi, 1984; Minami and Oshawa, 1990; Ronen et al., 1988; Schnabel and Stout, 1994; Terry et al., 1981). Transitorily significant quantities of N$_2$O can be dissolved in the soil solution. The maximal concentrations of dissolved N$_2$O reported are up to three orders of magnitude above ambient. On the other hand, temporary N$_2$O 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$_2$O in soils is very different. In some studies the concentrations of N$_2$O 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$_2$O concentrations in the soil profiles is in part the result of seasonal and chemical factors regulating N$_2$O dynamics. Some investigations revealed seasonal dynamics of N$_2$O concentrations in the soil solution (Bowden and Bormann, 1986; Minami and Fukushi, 1984). Other reports, however could not confirm such systematic relations (e.g. Minami and Oshawa, 1990).

Few investigations take the transport of dissolved N$_2$O with the soil solution and further out-gassing into the atmosphere. Leaching of dissolved N$_2$O requires high N$_2$O 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$_2$O production, the denitrification being a major N$_2$O production process, clear relationships between increasing concentrations of N$_2$O and decreasing availability of O$_2$ in the soil air have been often shown.
In wet soils the N$_2$O flux increased four times more slowly with rising N$_2$O 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 N$_2$O concentrations in the soil air and the N$_2$O 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 N$_2$O 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 N$_2$O (Goodroad and Keeney, 1985). Other factors may be large spatial variability of N$_2$O concentrations and N$_2$O fluxes or the dissolution of N$_2$O in the soil solution (Velthof et al., 1996). Furthermore, high water contents may restrict N$_2$O diffusion and enforce the microbial reduction of N$_2$O to N$_2$ (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 μmolCO$_2$ L$^{-1}$; 0.01–417 μmolCH$_4$ L$^{-1}$; 0–3.3 μmolN$_2$O L$^{-1}$. Pangala et al. (2013) reported the amount of CH$_4$ in deeper peat in Borneo peatland was greater (113–1539 μmol CH$_4$ 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 CH$_4$ concentrations of 175–1380 μmolCH$_4$ 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) O$_2$ 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 CH$_4$ and N$_2$O (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 O$_2$.

Though O$_2$ has long been recognized as an important driver for determining microbial metabolism, relatively little is known about field-level O$_2$ dynamics in soils (Burgin et al., 2010; Liptzin et al., 2011; Silver et al., 1999). Consequently, our understanding of how soil O$_2$ varies is limited. This restricts our understanding of how soil O$_2$ 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 O$_2$ 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 O$_2$, 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 (Kozlowski, 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 CO$_2$, CH$_4$, and N$_2$O in wetland ecosystems in the world.

<table>
<thead>
<tr>
<th>No.</th>
<th>Location</th>
<th>Land use</th>
<th>Peatland/soil type</th>
<th>Sampling period</th>
<th>Flux (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$^{-1}$))</th>
<th>Dissolved concentration</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Alpine wetland ecosystem in Luanhaizi, Qinghai-Tibetan Plateau, China 37° 29′ N, 101° 12′ E</td>
<td>Alpine meadow ecosystem Emergent-plant</td>
<td>July 2002 to September 2002</td>
<td>Mean 2.46</td>
<td>Mean 0.1 - 2.5 Mean 0.1 - 0.6</td>
<td></td>
<td>Hirota et al., 2004.</td>
</tr>
<tr>
<td>2</td>
<td>Peatland near St. Charles-de-Bellechasse, Quebec, Canada 46° 40′ N, 71° 10′ W</td>
<td>Hummock, lawn, and hollow forest</td>
<td>May 2002 and January 2003</td>
<td>Drained site Mean 1.4 to 5.8 (2002)</td>
<td>Mean 0.3 to 0.7 (2002)</td>
<td></td>
<td>Strack et al., 2004.</td>
</tr>
<tr>
<td>3</td>
<td>Peatland in the Mer Bleue, near Ottawa in eastern Ontario, Canada</td>
<td>Vegetations of mosses Oligotrophic peatland</td>
<td>June 2003 to December 2004</td>
<td>0.1-2.5</td>
<td>0.01-0.190</td>
<td></td>
<td>Blodau et al., 2007.</td>
</tr>
<tr>
<td>4</td>
<td>The Mer Bleue site near Ottawa, eastern Ontario, Canada</td>
<td>Mosses and shrubs Oligotrophic peatland</td>
<td>June 2003 to December 2004</td>
<td>0.97-6.5</td>
<td>0.05-1.2</td>
<td></td>
<td>Blodau and Moore, 2003.</td>
</tr>
<tr>
<td>7</td>
<td>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</td>
<td>Vegetation varies: Carex lasiocarpa Carex meyeriana Deyeuxia angustifolia</td>
<td>August 2001</td>
<td>17.2-45.0</td>
<td>8-202</td>
<td></td>
<td>Ding et al., 2004.</td>
</tr>
<tr>
<td>Table 2.1 continued</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>----------------------</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>8</strong> The Wildmooswald in the cool-humid Central Black Forest in Germany 47° 57′ N, 8° 07′ E</td>
<td>Mature Norway spruce</td>
<td>Haplic Gleysol</td>
<td>47-129</td>
<td>−0.024-0.072</td>
<td>81-449</td>
<td>Jungkunst <em>et al.</em>, 2008.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(Picea abies L.)</td>
<td>Histic Gleyso</td>
<td>60-158</td>
<td>−0.012-0.094</td>
<td>83-963</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sapric Histosol</td>
<td>65-172</td>
<td>0.010-0.141</td>
<td>20-653</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>9</strong> The north of Great Hing’an Mountains, northeast China 52.94° N, 122.86° E</td>
<td>Vascular plants</td>
<td>Minerotrophic peatland</td>
<td>June 2010 to September 2011</td>
<td>0.21-1.02</td>
<td>Mean 0.01437</td>
<td>Miao <em>et al.</em>, 2012.</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>10</strong> The Asa experimental forest, in southern Sweden 57° 08′ N, 14° 45′ E</td>
<td>Drained site and undrained site</td>
<td>Sedge mire</td>
<td>August 1999 to September 2002</td>
<td>Median 70 to 210</td>
<td>Median −5 to 1200</td>
<td>Median 1 to 4</td>
<td>Von Arnold <em>et al.</em>, 2005.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Lakes</td>
<td>Eutrophic to oligotrophic</td>
<td>June to July (Summer) 2009</td>
<td>−6.0-123.9</td>
<td>1.4-18.1</td>
<td>Schier-Uijl <em>et al.</em>, 2011.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Ditches</td>
<td></td>
<td></td>
<td>69.6-199.0</td>
<td>1.2-39.3</td>
<td></td>
</tr>
<tr>
<td><strong>11</strong> The east and southwest of the Netherlands</td>
<td>Pasture farm (mixed arable/grassland) Field drain Open farm yard ditch</td>
<td>Podzols, brown forest soil, and noncalcareous gleys soil</td>
<td>August 29, 2002</td>
<td></td>
<td></td>
<td>Reay <em>et al.</em>, 2004.</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Mean 28.2</td>
<td>Mean 8.5</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Mean 3.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>12</strong> Drainage catchment area, near the headwater of a tributary of Ythan River catchment, Aberdeenshire, UK</td>
<td>Combination of peat extraction: Bare peat Nonfertilized Phalaris site Fertilized Phalaris site Natural raised bog site</td>
<td>Mineralized fibric histosol</td>
<td>May 2010 to May 2011</td>
<td>0.0015-0.521</td>
<td>−0.021-0.9</td>
<td>−19-261</td>
<td>Mander <em>et al.</em>, 2012.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Table 2.1 continued</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>---------------------</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>No.</th>
<th>Location</th>
<th>Vegetation Type</th>
<th>Condition</th>
<th>Start Date - End Date</th>
<th>Value Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>15</td>
<td>Peat swamp basin in Mukah</td>
<td>Forest, Sago</td>
<td>November</td>
<td>2002-2003</td>
<td>63-245</td>
</tr>
<tr>
<td></td>
<td>Division of Sarawak</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Malaysia 2° 49' N, 111° 51' E</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>E to 2° 49' N, 111°56' E</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Oil palm</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>16</td>
<td>Peat swamp basin in Mukah</td>
<td>Forest, Sago</td>
<td>November</td>
<td>2002-2003</td>
<td>-0.000453-</td>
</tr>
<tr>
<td></td>
<td>Division of Sarawak</td>
<td></td>
<td></td>
<td></td>
<td>0.0084</td>
</tr>
<tr>
<td></td>
<td>Malaysia 2° 49' N, 111° 51' E</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>E to 2° 49' N, 111°56' E</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Oil palm</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>17</td>
<td>Peat swamp basin in Mukah</td>
<td>Forest, Sago</td>
<td>November</td>
<td>2002-2003</td>
<td>-3.4-19.7</td>
</tr>
<tr>
<td></td>
<td>Division of Sarawak</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Malaysia 2° 49' N, 111° 51' E</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>E to 2° 49' N, 111°56' E</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Oil palm</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>18</td>
<td>Tropical peatland, upper</td>
<td>Forest, Hollows</td>
<td>in 1999,</td>
<td>0.0329</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sebangau River,</td>
<td></td>
<td>2000, and</td>
<td>(50 - 150 cm depths)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Central Kalimantan</td>
<td></td>
<td>2001</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>0°20’ S, 113° 55' E</td>
<td></td>
<td>during wet</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>and dry</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>season</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Mean 139</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Maximum</td>
<td>0.35</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Mean 484</td>
<td></td>
<td></td>
</tr>
<tr>
<td>19</td>
<td>Tropical peatland, upper</td>
<td>Forest, Hollows</td>
<td>March 2011</td>
<td>Mean 0.0329</td>
<td>0.113-1.539</td>
</tr>
<tr>
<td></td>
<td>Sebangau River,</td>
<td></td>
<td></td>
<td>(50 - 150 cm depths)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Southeast of Palangka Raya,</td>
<td></td>
<td></td>
<td>Mean 0.007</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Central Kalimantan</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Stem of plants</td>
<td>0.017-0.185</td>
</tr>
</tbody>
</table>
### Table 2.1 continued

<table>
<thead>
<tr>
<th>Location</th>
<th>Type</th>
<th>Date</th>
<th>Description</th>
<th>Mean (all depths)</th>
<th>Mean (all depths)</th>
<th>Mean (all depths)</th>
<th>Mean (all depths)</th>
<th>Present study</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tropical peatland near Palangka Raya city, Central Kalimantan, Indonesia 02° 19′ 25.20″ S, 113° 54′ 16.86″ E</td>
<td>Forest sites: Typic Tropofibrists</td>
<td>December 2011 to May 2012</td>
<td>FW1</td>
<td>Mean 186</td>
<td>Mean 0.424</td>
<td>Mean −1.24</td>
<td>Mean 0.698</td>
<td>Present study</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>FW2</td>
<td>Mean 200</td>
<td>Mean 1.93</td>
<td>Mean −3.1</td>
<td>Mean 0.659</td>
<td>Present study</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>FD</td>
<td>Mean 340</td>
<td>Mean 0.008</td>
<td>Mean 3.4</td>
<td>Mean 2.03</td>
<td>Present study</td>
</tr>
<tr>
<td></td>
<td>Burnt sites:</td>
<td></td>
<td>BW1</td>
<td>Mean 353</td>
<td>Mean 8.32</td>
<td>Mean −1.5</td>
<td>Mean 1.24</td>
<td>Present study</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>BW2</td>
<td>Mean 120</td>
<td>Mean 4.46</td>
<td>Mean −11.1</td>
<td>Mean 1.10</td>
<td>Present study</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>BD</td>
<td>Mean 108</td>
<td>Mean 0.220</td>
<td>Mean 8.1</td>
<td>Mean 0.836</td>
<td>Present study</td>
</tr>
<tr>
<td>Kalampangan village near Palangka Raya, Central Kalimantan, Indonesia 2° S, 114° E</td>
<td>Natural forest</td>
<td>March 2002 to March 2004</td>
<td>Regenerated forest</td>
<td>Mean 5.0 (dry season) and 49 (wet season)</td>
<td>Takakai et al., 2006.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Burnt forest</td>
<td>Mean 19 (dry season) and 55 (wet season)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Grassland</td>
<td>Mean 8.5 (dry season) and 26 (wet season)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Agriculture:</td>
<td>Plot A</td>
<td>Mean 163 (dry season) and 4298 (wet season)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Takakai et al., 2006.
### Table 2.1 continued

<table>
<thead>
<tr>
<th>Plot</th>
<th>Mean value</th>
<th>Location</th>
<th>Methodology</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Plot B</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Plot C</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>22</td>
<td>Peatland at Marchantaria Island, an island in Amazon River, near Manaus, Brazil 03° 13.947' S, 59° 56.765' W</td>
<td>Shallow part of lake Camaleão (Floodplain ecosystem)</td>
<td>October to November 1996 and 1997</td>
</tr>
<tr>
<td>23</td>
<td>The Daeng and Bacho swamp peatland near Narathiwat province in Thailand</td>
<td>Mangrove forest</td>
<td>Transition of the sandy plains and the swamps</td>
</tr>
<tr>
<td>23</td>
<td>Peatland in southernmost Patagonia, Tierra del Fuego 54° 45' S, 68° 20' W</td>
<td>Vascular plants</td>
<td>Bog peatland</td>
</tr>
<tr>
<td>24</td>
<td>Wolong Marsh and Tuanjie Marsh in the north of Milor Peninsula of the Larseman Hills, eastern Arturica 69° 25' to 69° 56' S, 76° 20' to 76° 45' E.</td>
<td>Vegetations of algae community</td>
<td>Marsh (Wetland ecosystem)</td>
</tr>
</tbody>
</table>


Ueda et al., 2000.

Fritz et al., 2011.

Zhu et al., 2008.
Chapter 3
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.](image)

Annual precipitations and annual mean air temperature for the nine years of 2002–2010 were 2540±596 mm and 26.2±0.3°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 *Tristaniawhittiana*, *Combretocarpusrotundatus*, *Palaquiumleiocarpum*, and *Stemonurusscorpioides*. 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 (*Combretocarpusrotundatus*), 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±17.3 kg m$^{-3}$ (Shimada et al., 2001).
Chapter 4
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.
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:

\[ C_n = C^* a + C_{n-1}(1 - a) \]  \hspace{1cm} (4.1)

\[ C^* = \frac{[C_n - C_{n-1}(1 - a)]}{a} \] \hspace{1cm} (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<\( a <1 \)).
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).
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.
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).

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

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.
4.3.3 DO concentration

DO concentrations ranged from 0.5 to 25 μmol L$^{-1}$, which is generally <10 μmol L$^{-1}$, much lower than DO in a solution which is in equilibrium with the atmospheric O$_2$ level (220–280 μmol 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±2.9 μmol 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 μmol 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).

![Figure 4.8](image1.png)

**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).

![Figure 4.9](image2.png)

**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.

<table>
<thead>
<tr>
<th>Drainage condition</th>
<th>Depth</th>
<th>Land use</th>
<th>Natural forest</th>
<th>Burnt area</th>
<th>All</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Flooded condition</td>
<td>10 cm</td>
<td>5.63 ± 2.47 (7)</td>
<td>3.56 ± 1.33 (8)</td>
<td>4.53 ± 2.16 (15)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>20 cm</td>
<td>7.47 ± 6.61 (8)</td>
<td>3.98 ± 2.36 (8)</td>
<td>5.72 ± 5.12 (16)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>40 cm</td>
<td>7.27 ± 7.51 (8)</td>
<td>3.85 ± 1.78 (8)</td>
<td>5.56 ± 5.56 (16)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>80 cm</td>
<td>9.63 ± 7.79 (8)</td>
<td>5.19 ± 2.75 (7)</td>
<td>7.56 ± 6.23 (15)</td>
<td></td>
</tr>
<tr>
<td>All</td>
<td></td>
<td>7.56 ± 6.38 (31)</td>
<td>4.11 ± 2.08 (31)</td>
<td>5.84 ± 5.02 (62)</td>
<td></td>
</tr>
<tr>
<td>Drained condition</td>
<td>10 cm</td>
<td>6.92 ± 0.00 (1)</td>
<td>8.11 ± 8.47 (3)</td>
<td>7.81 ± 6.94 (4)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>20 cm</td>
<td>7.05 ± 2.43 (3)</td>
<td>2.20 ± 1.91 (4)</td>
<td>4.28 ± 3.25 (7)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>40 cm</td>
<td>5.05 ± 1.90 (4)</td>
<td>2.99 ± 1.94 (4)</td>
<td>4.02 ± 2.09 (8)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>80 cm</td>
<td>3.81 ± 1.06 (4)</td>
<td>3.18 ± 1.85 (4)</td>
<td>3.49 ± 1.44 (8)</td>
<td></td>
</tr>
<tr>
<td>All</td>
<td></td>
<td>5.29 ± 2.07 (12)</td>
<td>3.85 ± 4.19 (15)</td>
<td>4.49 ± 3.44 (27)</td>
<td></td>
</tr>
</tbody>
</table>

1. Values are means and standard deviations of DO concentration in subsurface water (μmol 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.

<table>
<thead>
<tr>
<th>DO</th>
<th>df</th>
<th>F</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Main factors</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Land use</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Drainage</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Depth</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Error</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Interactions</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Land use*Drainage</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Land use*Depth</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Drainage*Depth</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Land use<em>Drainage</em>Depth</td>
</tr>
</tbody>
</table>

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 –COOH, phenol-OH, alcohol-OH, -NH2 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](image-url) 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 μS 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).

<table>
<thead>
<tr>
<th>Site</th>
<th>Description</th>
<th>pH</th>
<th>EC  (µS cm⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>FW1</td>
<td>Flooded forest site #1</td>
<td>3.6 ± 0.1 ab (31)</td>
<td>85 ± 8 c (31)</td>
</tr>
<tr>
<td>FW2</td>
<td>Flooded forest site #2</td>
<td>3.6 ± 0.2 ab (31)</td>
<td>86 ± 14 c (31)</td>
</tr>
<tr>
<td>FD</td>
<td>Drained forest site</td>
<td>3.5 ± 0.1 a (17)</td>
<td>125 ± 16 d (17)</td>
</tr>
<tr>
<td>BW1</td>
<td>Flooded burnt site #1</td>
<td>3.7 ± 0.2 b (31)</td>
<td>56 ± 10 b (31)</td>
</tr>
<tr>
<td>BW2</td>
<td>Flooded burnt site #2</td>
<td>3.7 ± 0.2 b (32)</td>
<td>50 ± 12 b (32)</td>
</tr>
<tr>
<td>BD</td>
<td>Drained burnt site</td>
<td>4.0 ± 0.2 c (28)</td>
<td>35 ± 5 a (28)</td>
</tr>
</tbody>
</table>

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

NO₃⁻-N concentration ranged from 0.0519 to 0.260 mg L⁻¹. The mean of NO₃⁻-N concentration was generally higher in the burnt sites than in the forest sites. NO₃⁻-N was highest at BW1, followed by BW2>BD>FD>FW1>FW2 (Table 4.5). There was significant difference only between BW2 and the flooded forest sites (FW1 and FW2). At BW1 and BW2, NO₃⁻-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).

<table>
<thead>
<tr>
<th>Site</th>
<th>Description</th>
<th>NO$_3^-$ -N (mg L$^{-1}$)</th>
<th>NH$_4^+$ -N (mg L$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>FW1</td>
<td>Flooded forest site #1</td>
<td>0.0565 ± 0.0492 a (15)</td>
<td>0.0741 ± 0.060928 (15)</td>
</tr>
<tr>
<td>FW2</td>
<td>Flooded forest site #2</td>
<td>0.0519 ± 0.0319 a (13)</td>
<td>0.143 ± 0.214 (13)</td>
</tr>
<tr>
<td>FD</td>
<td>Drained forest site</td>
<td>0.0781 ± 0.0325 ab (9)</td>
<td>0.086 ± 0.130 (9)</td>
</tr>
<tr>
<td>BW1</td>
<td>Flooded burnt site #1</td>
<td>0.260 ± 0.253 b (15)</td>
<td>0.194 ± 0.491 (15)</td>
</tr>
<tr>
<td>BW2</td>
<td>Flooded burnt site #2</td>
<td>0.211 ± 0.219 ab (15)</td>
<td>0.205 ± 0.263 (15)</td>
</tr>
<tr>
<td>BD</td>
<td>Drained burnt site</td>
<td>0.0915 ± 0.0401 ab (13)</td>
<td>0.061 ± 0.110 (13)</td>
</tr>
</tbody>
</table>

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$_2$O. Thus the effects of NH$_4^+$ and NO$_3^-$ are discussed in relation to the flux and dissolved concentration of N$_2$O 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, $6.9\pm5.6$ and $4.0\pm2.9$ μmol 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.
Chapter 5
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 ($\Delta C_c$) against closure time ($\Delta t$):

$$F = \rho_g \times (V/A) \times (\Delta C_c / \Delta t) \times [273/(273 + T)]$$  \hspace{1cm} (5.1)

where $F$ is the gas flux (gC or gN m$^{-2}$ h$^{-1}$), $\rho_g$ is the gas density (0.536 × 10$^3$ g m$^{-3}$ for CO$_2$-C and CH$_4$-C; 1.259 × 10$^3$ g m$^{-3}$ for N$_2$O-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 CO$_2$-equivalent carbon flux (gCO$_2$-eq C m$^{-2}$ h$^{-1}$) by the following equation:


\[ GWP = F_{CO2} + F_{CH4} \times \frac{16}{12} \times 25 \times \frac{12}{44} + F_{N2O} \times \frac{44}{28} \times 298 \times \frac{12}{44} \]  \hspace{1cm} (5.2)

where 25 and 298 are the factors for converting CH₄ and N₂O fluxes to their GWP 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 mgC m⁻² h⁻¹ (Table 5.1). The high CH₄ emissions at BW1 and BW2 and negative CH₄ fluxes (= CH₄ absorption) at the forest sites were periodically observed (Fig. 5.2). At FD site, 7 of 15 measured fluxes were negative. However, on average, all sites were a net CH₄ sources during the observation period (Table 5.1). In the burnt sites, higher average CH₄ fluxes were associated with the higher average water table levels (Fig. 5.3). Relative to the burnt sites, the CH₄ fluxes at the forest sites were consistently lower regardless of water table levels. The results of two-way ANOVA indicated that both main effects (land use and drainage) and their interaction were significant, and CH₄ flux at the flooded burnt sites was significantly higher than those of the other sites (Table 5.4). This means the CH₄ flux under flooded conditions was significant higher than under drained conditions in both land use during the field observation.
Fig. 5.2  Seasonal variations of CH$_4$ fluxes measure at the ground surface in the study sites. Error bars show standard deviation.

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

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

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 mgC m⁻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 CO$_2$ fluxes measured at the ground surface in the study sites. Error bars show standard deviation.

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

<table>
<thead>
<tr>
<th>Drainage</th>
<th>Land use</th>
<th>CO₂ flux (mgC m(^{-2}) h(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Natural forest</td>
<td>Burnt area</td>
</tr>
<tr>
<td>Flooded</td>
<td>195 ± 199 (23)</td>
<td>198 ± 165 (18)</td>
</tr>
<tr>
<td>Drained</td>
<td>340 ± 250 (15)</td>
<td>108 ± 115 (15)</td>
</tr>
<tr>
<td>All</td>
<td>252 ± 229 (38)</td>
<td>157 ± 150 (33)</td>
</tr>
</tbody>
</table>

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 N\(_2\)O flux

The site-specific averages of N\(_2\)O flux ranged from −8.7±41.9 to 8.1±75.5 μgN m\(^{-2}\) h\(^{-1}\) (Table 5.3). Although occasional variations were observed in early Apr. at BD and BW2, little variation was observed in N\(_2\)O flux at other sites during the observation period (Fig. 5.6). On average, the N\(_2\)O 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 N\(_2\)O in tropical peatlands. However, the relationship between N\(_2\)O flux and water table level was unclear (Fig. 5.7). The results of two-way ANOVA suggested that there were no significant differences in N\(_2\)O fluxes between sites based on either land use or drainage (Table 5.4). The N\(_2\)O flux measured at each chamber at study sites ranged from −155 to 218 μgN m\(^{-2}\) h\(^{-1}\).
Fig. 5.6  Seasonal variations of N$_2$O fluxes measure at the ground surface in the study sites. Error bars show standard deviation.

Fig. 5.7  Relationship between N$_2$O 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.

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

1. Values are means and standard deviations of N₂O flux (µgN m⁻² 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.

<table>
<thead>
<tr>
<th>Sites</th>
<th>CH$_4$ flux (mgC m$^{-2}$ h$^{-1}$)</th>
<th>CO$_2$ flux (mgC m$^{-2}$ h$^{-1}$)</th>
<th>N$_2$O flux (μgN m$^{-2}$ h$^{-1}$)</th>
<th>GWP (CO$_2$-eq mgC m$^{-2}$ h$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Flooded forest sites (FW1 and FW2)</td>
<td>1.37 ± 2.03 a (24)</td>
<td>195 ± 199 ab (23)</td>
<td>-2.4 ± 16.9 (24)</td>
<td>208 ± 204 ab (23)</td>
</tr>
<tr>
<td>Drained forest site (FD)</td>
<td>0.0084 ± 0.0321 a (15)</td>
<td>340 ± 250 b (15)</td>
<td>3.4 ± 19.2 (15)</td>
<td>340 ± 250 b (15)</td>
</tr>
<tr>
<td>Flooded burnt sites (BW1 and BW2)</td>
<td>5.75 ± 6.66 b (18)</td>
<td>198 ± 165 ab (18)</td>
<td>-8.7 ± 41.9 (16)</td>
<td>249 ± 171 ab (18)</td>
</tr>
<tr>
<td>Drained burnt site (BD)</td>
<td>0.220 ± 0.143 a (15)</td>
<td>108 ± 115 a (15)</td>
<td>8.1 ± 75.5 (14)</td>
<td>111 ± 113 a (15)</td>
</tr>
</tbody>
</table>

**ANOVA**

<table>
<thead>
<tr>
<th></th>
<th>df</th>
<th>F</th>
<th>df</th>
<th>F</th>
<th>df</th>
<th>F</th>
<th>df</th>
<th>F</th>
</tr>
</thead>
<tbody>
<tr>
<td>Main factors and interaction</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Land use</td>
<td>1</td>
<td>7.34 **</td>
<td>1</td>
<td>6.33 *</td>
<td>1</td>
<td>0.01</td>
<td>1</td>
<td>4.13 *</td>
</tr>
<tr>
<td>Drainage</td>
<td>1</td>
<td>16.49 **</td>
<td>1</td>
<td>0.36</td>
<td>1</td>
<td>1.23</td>
<td>1</td>
<td>0.00</td>
</tr>
<tr>
<td>Land use*Drainage</td>
<td>1</td>
<td>6.05 *</td>
<td>1</td>
<td>6.64 *</td>
<td>1</td>
<td>0.29</td>
<td>1</td>
<td>8.59 **</td>
</tr>
<tr>
<td>Error</td>
<td>68</td>
<td>67</td>
<td>65</td>
<td>67</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Simple main effects**

<table>
<thead>
<tr>
<th></th>
<th>df</th>
<th>F</th>
<th>df</th>
<th>F</th>
<th>df</th>
<th>F</th>
<th>df</th>
<th>F</th>
</tr>
</thead>
<tbody>
<tr>
<td>Land use under flooded condition</td>
<td>15.85 **</td>
<td>0.00</td>
<td>0.22</td>
<td>0.47</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Land use under drained condition</td>
<td>0.03</td>
<td>11.30 **</td>
<td>0.09</td>
<td>10.73 **</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Drainage in forest vegetation</td>
<td>1.36</td>
<td>5.32 *</td>
<td>0.18</td>
<td>4.34 *</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Drainage in burnt area</td>
<td>20.05 **</td>
<td>1.86</td>
<td>1.23</td>
<td>4.26 *</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

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](image-url) 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>
<thead>
<tr>
<th>Drainage</th>
<th>Land use</th>
<th>Natural forest</th>
<th>Burnt area</th>
<th>All</th>
</tr>
</thead>
<tbody>
<tr>
<td>Flooded</td>
<td></td>
<td>208 ± 204 (23)</td>
<td>249 ± 171 (18)</td>
<td>226 ± 189 (41)</td>
</tr>
<tr>
<td>Drained</td>
<td></td>
<td>340 ± 250 (15)</td>
<td>111 ± 113 (15)</td>
<td>226 ± 224 (30)</td>
</tr>
<tr>
<td>All</td>
<td></td>
<td>260 ± 230 (38)</td>
<td>186 ± 161 (33)</td>
<td>226 ± 203 (71)</td>
</tr>
</tbody>
</table>

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 CH\textsubscript{4} flux in tropical peatlands

5.4.1.1 Effect of water table level

In this study, especially in burnt sites, CH\textsubscript{4} 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 CH\textsubscript{4} 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 O\textsubscript{2} to support methanotrophy (Bartlett et al., 1988; Koschorreck, 2000).

Based on a comprehensive review on CH\textsubscript{4} flux observation in tropical peatlands, Couwenberg et al. (2010) has recently reported that CH\textsubscript{4} 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 CH\textsubscript{4} 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 CH\textsubscript{4} 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 CH\textsubscript{4} fluxes ranged from −4.53 to 8.40 μgC m\textsuperscript{2} h\textsuperscript{−1} in a mixed peat swamp forest (water table level: −60 to −20 cm), from −7.44 to 102 μgC m\textsuperscript{2} h\textsuperscript{−1} in a sago plantation (−40 to 0 cm), and from −32.8 to 4.17 μgC m\textsuperscript{2} h\textsuperscript{−1} in an oil palm plantation (−80 to −30 cm) and also increased with increasing water table levels.

In drained forest site, in contrast, CH\textsubscript{4} 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 CH\textsubscript{4} 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 O\textsubscript{2} 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).
CH$_4$ fluxes and trends in the drained sites were similar to those reported in these previous studies (Table 5.4). Considering the relationship between CH$_4$ flux and water table level cited above, there is an empirical depth of water table level at which CH$_4$ production and oxidations are balanced. In this study, the drained forest site was both a weak sink and source of CH$_4$ (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 CH$_4$ from tropical peatlands is complex. Aerenchyma promoted gas exchange between the atmosphere and the rhizosphere (Megonigal and Day, 1992; Kozlowski, 1997), in particular entry of O$_2$ to the root zone. The rhizosphere is often a site of major microbiological activity (Curl and Truelove, 1986) and subsurface emission of O$_2$ 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 CH$_4$ emissions from wetlands, because two gas-transport mechanisms, molecular diffusion and convective gas flow, are well documented in some plants (Brix et al., 1992; Nouchi and Mariko, 1993). In addition, near-surface O$_2$ level is affected by transport linked to the presence of plants and O$_2$ release from plant roots (Elberling et al., 2011).

CH$_4$ 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 CH$_4$ produced in the peat might be oxidized by plant-mediated oxygen supply in the forest sites. Blodau and Moore (2003) observed a depletion of O$_2$ and initiation of CH$_4$ 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 respiration 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 N$_2$O into peat water for denitrification is also related to the amount of NO$_3^-$ (Burgin and Groffman, 2012). A recent review indicated that net negative N$_2$O fluxes have been reported in numerous previous studies, showing that low mineral nitrogen and high moisture content are favorable for N$_2$O consumption (Chapuis-Lardy et al., 2007). NO$_3^-$ 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 N$_2$O 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 N$_2$O emitted into the atmosphere as it contributes approximately 60% of the emissions globally (Ehhalt et al., 2001). The application of N fertilizer generally increases N$_2$O emission rate from the crop land. The increasing use of inorganic N fertilizers under humid tropical climate may contribute to N$_2$O 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 N$_2$O emission.

NO$_3^-$ concentration is one of the important controlling factors for soil N processes, including nitrification and denitrification. In this study, the effect of NO$_3^-$ on N$_2$O flux was unclear, although there were significant differences in dissolved NO$_3^-$ concentrations in peat water among the study sites (Table 4.5). Takakai et al. (2006) found that the N$_2$O flux in a cropland in the vicinity of our study area increased with increasing NO$_3^-$-N content in the top 10 cm of soil during rainy season. In that study, however, the NO$_3^-$-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 NO$_3^-$-N content was likely too small to influence N$_2$O emissions. In addition, Inubushi et al. (2003) have investigated seasonal changes in N$_2$O emission in N$_2$O emission over a whole year in tropical peatland in South Kalimantan. N$_2$O 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 N$_2$O emission. They explained this result as inhibition of N$_2$O emission by flood water.
5.4.4 Effect of each GHG flux on GWP

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

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

5.5 Summary

The investigations of field study showed that: 1) CH₄ fluxes were significantly influenced by land use and drainage, which were highest in the flooded burnt sites (5.75±6.66 mgC m⁻² h⁻¹) followed by the flooded forest sites (1.37±2.03 mgC m⁻² h⁻¹), the drained burnt site (0.220±0.143 mgC m⁻² h⁻¹), and the drained forest site (0.0084±0.0321 mgC m⁻² h⁻¹), 2) CO₂ fluxes were significantly higher in the drained forest site (340±250 mgC m⁻² h⁻¹ with the water table level of −20 to −60 cm) than in the drained burnt site (108±115 mgC m⁻² h⁻¹ with the water table level of −15 to +10 cm), and 3) N₂O fluxes ranged from −2.4 to −8.7 μgN m⁻² h⁻¹ in the flooded sites and from 3.4 to 8.1 μgN m⁻² h⁻¹ in the drained sites. The negative N₂O fluxes might be caused by N₂O consumption by denitrification under flooded conditions. GWP was mainly determined by CO₂ flux, with highest levels in the drained forest site.
Chapter 6
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 CH$_4$ and N$_2$O and in-situ equilibration chamber for CO$_2$.

6.2.1 Concentrations of dissolved CH$_4$ and N$_2$O
Gas samples for the analyses of dissolved CH$_4$ and N$_2$O 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 O$_2$ 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 O$_2$-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 CH$_4$ and N$_2$O in the bottles were determined in the same manner used for flux measurements for these gases.
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$_2$O 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.
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 ($C_t$, mol mol⁻¹) in this method (dissolved CH₄, CO₂, and N₂O concentrations) was given by Henry’s law:

$$
C_t = \frac{C_g \times 10^{-6} \times P}{K_H}
$$

(6.1)

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

---

**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.
differences of ±4.2% in estimation of dissolved CO₂ by Eq. 6.1. The Cᵢ in molar ratio was converted to Cᵢ (mol L⁻¹) according to the following equation:

\[ Cᵢ [\text{mol L}^{-1}] = Cᵢ [\text{mol mol}^{-1}] \times \rho_w \]

(6.2)

where \( \rho_w \) is the water density (1000 g L⁻¹), and \( m_w \) is the molecular weight of water (H₂O, 18.01 g mol⁻¹).

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

\[ Cᵢ = \frac{C_g (V_g + \alpha \cdot V_l) - C_{bg} \cdot V_g}{V_l} \]

(6.3)

where \( C_{bg} \) is the initial concentration of the gas in the O₂-free air, \( \alpha \) 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) \( V_g = V_l \):

\[ Cᵢ = C_g (1 + \alpha) \]

(6.4)

Here, \( \alpha \) is defined as \( C'ᵢ/C_g \), where \( C'ᵢ \) is the dissolved gas concentration in the liquid phase inside the syringe. The variable \( C'ᵢ \) was calculated in the same manner as \( Cᵢ \) 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 \( C'ᵢ \) as follows:

\[ C_g [\text{mol L}^{-1}] = \frac{C_g [\text{ppmv}] \times 10^{-6}}{V} \]

(6.5)

where \( V \) (L mol⁻¹) is the volume of 1-mol air at total pressure \( P \) (= 1.0 atm) and temperature \( T \) (°C), can be calculated as:

\[ V = 22.4 \times \frac{1}{P} \times \frac{T + 273}{273} \]

(6.6)

6.2.4 Statistical analysis

The effects of land use, drainage, and depth on dissolved GHG concentrations were tested using three-way ANOVA. Multiple comparison tests among the sites and depths were conducted using the Bonferroni method. A probability level less than 0.05 (\( P<0.05 \)) was
considered statistically significant. All analyses were performed in IBM SPSS Statistics Ver. 21.

6.3 Results

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

6.3.1 Dissolved CH₄ concentration

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

The result of the statistical analysis (Table 6.4) shows that dissolved CH₄ concentration significantly affected by all main factors, namely land use, drainage, and depth (P<0.01 in all cases), meanwhile all the two-way interactions were significant (P<0.01 in all cases). The results of the post hoc test were as follows. The effect of land use was significant at all depths under flooded condition (P<0.001, forest<burnt). Under the drained condition, dissolved CH₄ concentration at 80 cm was nearly significant (P=0.059, forest<burnt). The effect of drainage was significant only in the burnt sites at 20, 40, and 80 cm (P<0.001 in all cases; flooded>drained). The effect of depth was significant only in the drained burnt site (P<0.001; 20 cm>others depths). In short, dissolved CH₄ concentrations in the flooded burnt sites was significantly higher than those in the other sites, especially much higher at 20 cm. Dissolved
CH$_4$ in the drained burnt sites also significantly higher than those in the forest sites except for 20 cm.

**Table 6.1** Dissolved CH$_4$ concentrations under different land use and drainage conditions averaged for the observation period.

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

1. Values are means and standard deviations of dissolved CH$_4$ 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 CH$_4$ concentrations in pore water averaged for depths of 20–80 cm at each site. 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.

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

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$_2$O 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$_2$O concentrations in pore water averaged for the observation period (Dec. 2011 to May 2012). Error bars show standard deviation.
**Table 6.3** Dissolved N$_2$O concentrations under different land use and drainage conditions averaged for the observation period.

<table>
<thead>
<tr>
<th>Drainage</th>
<th>Depth</th>
<th>Land use</th>
<th>Dissolved N$_2$O</th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Natural forest</td>
<td>Burnt area</td>
<td>All</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Flooded</td>
<td>20 cm</td>
<td>0.007 ± 0.005 (14)</td>
<td>0.037 ± 0.120 (14)</td>
<td>0.022 ± 0.085 (28)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>40 cm</td>
<td>0.005 ± 0.003 (14)</td>
<td>0.006 ± 0.003 (13)</td>
<td>0.005 ± 0.003 (27)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>60 cm</td>
<td>0.005 ± 0.004 (14)</td>
<td>0.005 ± 0.002 (14)</td>
<td>0.005 ± 0.003 (28)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>80 cm</td>
<td>0.005 ± 0.003 (14)</td>
<td>0.005 ± 0.001 (14)</td>
<td>0.005 ± 0.002 (28)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>All</td>
<td>0.005 ± 0.004 (56)</td>
<td>0.013 ± 0.061 (55)</td>
<td>0.009 ± 0.043 (111)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Drained</td>
<td>20 cm</td>
<td>0.15 ± 0.28 (7)</td>
<td>0.006 ± 0.002 (7)</td>
<td>0.079 ± 0.205 (14)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>40 cm</td>
<td>0.22 ± 0.51 (7)</td>
<td>0.005 ± 0.002 (7)</td>
<td>0.113 ± 0.361 (14)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>60 cm</td>
<td>0.018 ± 0.027 (7)</td>
<td>0.005 ± 0.003 (7)</td>
<td>0.012 ± 0.019 (14)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>80 cm</td>
<td>0.007 ± 0.004 (7)</td>
<td>0.006 ± 0.003 (7)</td>
<td>0.006 ± 0.003 (14)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>All</td>
<td>0.099 ± 0.288 (28)</td>
<td>0.006 ± 0.002 (28)</td>
<td>0.052 ± 0.207 (56)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

1. Values are means and standard deviations of dissolved N$_2$O 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.

<table>
<thead>
<tr>
<th></th>
<th>Dissolved CO$_2$</th>
<th>Dissolved CH$_4$</th>
<th>Dissolved N$_2$O</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>df</td>
<td>F</td>
<td>df</td>
</tr>
<tr>
<td><strong>Main factors</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Land use</td>
<td>1</td>
<td>8.54 **</td>
<td>1</td>
</tr>
<tr>
<td>Drainage</td>
<td>1</td>
<td>14.48 **</td>
<td>1</td>
</tr>
<tr>
<td>Depth</td>
<td>3</td>
<td>4.09 **</td>
<td>3</td>
</tr>
<tr>
<td>Error</td>
<td>137</td>
<td>151</td>
<td>151</td>
</tr>
<tr>
<td><strong>Interactions</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Land use*Drainage</td>
<td>1</td>
<td>47.98 **</td>
<td>1</td>
</tr>
<tr>
<td>Land use*Depth</td>
<td>3</td>
<td>6.70 **</td>
<td>3</td>
</tr>
<tr>
<td>Drainage*Depth</td>
<td>3</td>
<td>9.36 **</td>
<td>3</td>
</tr>
<tr>
<td>Land use<em>Drainage</em>Depth</td>
<td>3</td>
<td>1.67</td>
<td>3</td>
</tr>
</tbody>
</table>

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

6.4.1 Concentrations of dissolved CH$_4$ in tropical peatlands

In contrast to the CH$_4$ fluxes, dissolved CH$_4$ concentrations were mainly affected by land use (Fig. 6.5). Ueda et al. (2000) found that dissolved CH$_4$ concentrations in groundwater in a coastal peat swamp in Thailand varied from 0.01 to 417 μmol L$^{-1}$, with average values of 48 and 226 μmol L$^{-1}$ in wet seasons spanning 4 years at two sampling sites. Koschorreck (2000) reported that CH$_4$ in pore water at the top 8 cm of a silty loam sediment on an island in the Amazon River ranged from 0 to 900 μmol L$^{-1}$. Terazawa et al. (2007) found that dissolved CH$_4$ concentrations in groundwater ranged from 5.6 to 28.4 μmol L$^{-1}$ in a floodplain forest located in northern Japan. Pangala et al. (2013) recently reported that dissolved CH$_4$ concentrations observed in the vicinity of the flooded forest sites in this study ranged from 113 to 1539 μmol L$^{-1}$, much higher than our results. Several differences between the two studies, including sampling depths (50 to 150 cm in Pangala et al., 2013 compared to 20 to 80 cm in this study) and procedures may have resulted in this discrepancy in dissolved CH$_4$ concentrations. However, the primary reason for this difference is unclear.

In this study, dissolved CH$_4$ concentrations were highest in the flooded burnt sites (124±84 μmol L$^{-1}$) followed by the drained burnt site (45.2±29.8 μmol L$^{-1}$), the flooded forest sites (1.15±1.38 μmol L$^{-1}$), and the drained forest site (0.860±0.819 μmol L$^{-1}$). One of the possible reasons of such high dissolved CH$_4$ in burnt sites is the lack of large trees which have thick and deep root systems. This situation inhibited the entry of O$_2$ to the soil profiles. Previous studies have also reported increased dissolved CH$_4$ as well as decreased DO (Liebner et al., 2012), redox potential (Fritz et al., 2011; Koschorreck, 2000), and root density (Fritz et al., 2011) in saturated peat profiles.

DO concentrations were influenced by land use only (Table 4.3) and were significantly higher in the forest sites (6.9±5.6 μmol L$^{-1}$) compared to the burnt sites (4.0±2.9 μmol L$^{-1}$) (Table 4.2). This suggested that CH$_4$ produced in the peat might be oxidized by plant-mediated oxygen supply in the forest sites. O$_2$ supply via plant roots would reduce net CH$_4$ production in the forest sites even under flooded conditions. The decline in DO by CH$_4$ oxidation could be compensated by this plant-mediated O$_2$ supply. The increase in the concentration of dissolved CH$_4$ concentrations with depth in the flooded forest sites suggested that CH$_4$ oxidation rate by plant-derived O$_2$ may depend on the amount of plant root biomass, which usually decreases with increasing depth.
The CH$_4$ 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 CH$_4$, while well-drained soils act as a sink for atmospheric CH$_4$ due to CH$_4$ oxidation. In the drained burnt site, CH$_4$ fluxes were lower than in the flooded forest sites, although dissolved CH$_4$ concentrations were much higher in the burnt sites compared to the forest sites (Fig. 6.5). Previous studies suggested that CH$_4$ diffusing toward the atmosphere is oxidized to CO$_2$ 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 CH$_4$ produced in deeper peat layers to be oxidized and CH$_4$ emissions and dissolved CH$_4$ concentrations at a depth of 20 cm to decrease. However, in the flooded burnt sites, the lack of an aerobic surface layer prohibited CH$_4$ oxidation, resulting in a higher CH$_4$ flux compared to other sites and a higher dissolved CH$_4$ concentration at 20 cm compared to deeper layers. Nouchi and Mariko (1993) reported that CH$_4$ emission rate was proportional to pore water CH$_4$ concentration, so the large CH$_4$ emissions and high dissolved CH$_4$ concentrations were mutually consistent.

### 6.4.2 Concentrations of dissolved CO$_2$ in tropical peatlands

The dissolved CO$_2$ concentrations were significantly higher in the drained forest site at depths of 40, 60, and 80 cm. These results were correspondent with the enhanced CO$_2$ 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 CO$_2$ concentrations at 20 cm were lower than those at deeper layers (Table 6.2), which could be attributed to diffusive CO$_2$ 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 CO$_2$ concentrations in a coastal swamp system in Thailand varied widely in swamp water, river water, and groundwater: 1 to 3290 μmol CO$_2$ L$^{-1}$, similar to the range in this study. The dissolved CO$_2$ 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 CO\(_2\) production through peat decomposition would suppress CO\(_2\) emissions and result in lower dissolved CO\(_2\) concentrations. In the flooded burnt sites, a continuously flooded condition (Fig. 4.5) would inhibit CO\(_2\) diffusion into the atmosphere and cause the high dissolved CO\(_2\) concentrations at 20 cm.

6.4.3 Concentrations of dissolved N\(_2\)O in tropical peatlands

According to Ueda et al. (2000) the dissolved N\(_2\)O 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 O\(_2\) to soil profile, which is favorable for denitrification process. In the flooded sites, the denitrification is a major N\(_2\)O production process. Clear relationships between increasing concentrations of N\(_2\)O 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). Some studies found no or only weak positive correlations of the N\(_2\)O concentration in the soil air and the N\(_2\)O 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 N\(_2\)O concentration at the flooded sites was lower than the drained forest site. Flooding of soils usually enforces denitrification, but N\(_2\)O flux was negative (sink) in the flooded sites. This could be probably due to denitrification which is involving the reduction of N\(_2\)O to N\(_2\). Under saturated conditions, a temporary use of atmospheric N\(_2\)O as an electron acceptor by microorganism probably occurred under anoxic conditions. It was explained that at the higher water table level, almost no N\(_2\)O emitted since most of N\(_2\)O had been converted microbiologically into N\(_2\) during the transport (Kliewer and Gilliam, 1995).

Nitrification is also an important microbial N process that produces N\(_2\)O. In this study, dissolved N\(_2\)O was significantly higher in drained forest sites than other sites at depths of 20 and 40 cm. In addition, N\(_2\)O 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 N\(_2\)O 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.
Chapter 7
GENERAL DISCUSSION

7.1 Natural peat swamp forest and their contributions on C balance

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

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

As with all forest, forested wetlands (swamps) are long-term sinks of atmospheric CO₂ in the absence of significant disturbances (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: Elaocarpaceae (Elaeocarpus mastersii), Ebenaceae (Diospyros bantamensis), Myrtaceae (Tristaniopsis sp. 2), Clusiaceae or Guttiferae (Mesua sp. 1), Lauraceae (Litsea elliptica), and Annonaceae (Xylopiafusca) (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-Pschorr et al., 1986; Jia et al., 2001; Yagi et al., 1996). Besides transporting CH₄, plants can also stimulate CH₄ production by providing substrates for methanogens (Lai, 2009; Whalen, 2005) and facilitate CH₄ oxidation by delivering O₂ downward into the rhizosphere for methanotrophs (Jia et al., 2001; van der Nat and Midderberg, 1998), and also transport labile
carbon compound into anoxic soil layers (Chanton et al., 2008; Joabsson and Christensen, 2001; Ström et al., 2003).

The low N$_2$O fluxes and dissolved N$_2$O concentrations in these study sites were probably due to the N$_2$O 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 O$_2$ diffusion into the subsurface layer through the air-filled pore space.

The higher CO$_2$ 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 CO$_2$ fluxes in the drained forest site suggested that CO$_2$ 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 CO$_2$ emissions at this site. Increased CO$_2$ emissions at lower water table levels are attributed to increased O$_2$ diffusion into unsaturated peat (Clymo, 1983), aerobic respiration being more efficient in CO$_2$ production than anerobic respiration (Schlesinger, 1997), and increased transportation of CO$_2$ through unsaturated peat (Moore and Dalva, 1993).

In laboratory incubations, low water table treatments increased CO$_2$ 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 CO$_2$ emissions at lower water table levels (Chimner, 2000; Kim and Verma, 1992; Silvola et al., 1996). Increased CO$_2$ emissions at lower water table levels are attributed to increased O$_2$ diffusion into unsaturated peat (Clymo, 1983). When peatlands are drained,
aerobic activity becomes possible even in deeper layers and the increased decomposition of the peat will increase CO₂ fluxes to the atmosphere (Laine and Päivänen, 1992). Drainage to greater depths increases the area of profile oxidation and, thus, tends to increase C loss (Armentano and Menges, 1986; Furukawa et al., 2005).

In addition, peatland drainage for cultivation can even be net sinks of CH₄ due to oxidation of CH₄ to CO₂ in the upper aerated peat layers (Maljanen et al., 2003a). The availability of O₂ is important for heterotrophic and autotrophic respiration, as the rate in oxic zones is considerably higher than in anoxic zones (Öquist and Sundh, 1998; Updegraff et al., 1995). The low CH₄ fluxes in the drained sites (Table 5.1) can be attributed to their oxic conditions. Drainage and subsequent oxidation of wetland profiles can decrease CH₄ production (Wang et al., 1996). In addition, soil-atmosphere CH₄ exchange is the result of simultaneously occurring production and consumption processes in soils, and is thus controlled by CH₄-producing methanogens operating at aerobic conditions and CH₄-consuming methanotrophs that depend on O₂ as a terminal electron acceptor (Topp and Pattey, 1997). Water table exerts a major control on CH₄ flux from wetlands through its 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 (Crill 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 (Kozlowski, 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$_2$O flux
was unclear, probably because the NO$_3^-$ concentration was too small to influence N$_2$O
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$_2$O and it contributed increase N$_2$O 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.
Chapter 8

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 CH$_4$ dynamics at the vegetated and non-vegetated lands in tropical peatland, the following hypotheses are addressed: 1) DO supplied from plant roots should oxidize CH$_4$ produced in saturated tropical peat soil, and 2) this CH$_4$ oxidation should reduce total GWP in a tropical peat swamp forest. 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. The CH$_4$ flux and the dissolved CH$_4$ concentrations 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. N$_2$O 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 CH$_4$ production rates (sink). But on the other hand, drained peatland enhances aerobic decomposition process, which increased the CO$_2$ fluxes to the atmosphere (source). Thus, soil also becomes as a source for N$_2$O 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 CH$_4$ 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.
References


103


Conrad R 1996: Soil microorganisms as controllers of atmospheric trace gases (H\textsubscript{2}, CO, CH\textsubscript{4}, OCS, N\textsubscript{2}O, and NO). *Microbiological reviews*, 60, 609-640.


EPA (Environmental Protection Agency) 2010: Methane and nitrous oxide emissions from natural source. United States Environmental Protection Agency Office of Atmospheric Programs. Washington DC. USA.


Pangala SR, Moore S, Hornibrook ERC, Gauci V 2013: Tree are major conduits for methane egress from tropical forested wetlands. *New Phytologist, 197*, 524-531.


Rieley JO, Page SE (eds) 1997: Biodiversity and sustainability of tropical peatlands. Samara publishing Ltd, Cardigan, UK.


Sebacher DI, Harriss RC, Bartlett KB 1985: Methane emissions to the atmosphere through aquatic plants. Journal of Environmental Quality, 14, 40-46


Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL (eds) Climate change 2007: the physical science basis. Contribution of working group I to the fourth assessment report of the intergovernmental panel on climate change. Cambridge University press, Cambridge. UK.


Verwer C, van der Meer P 2010: Carbon pools in tropical peat forest: towards a reference value for forest biomass in relatively undisturbed peat swamp forest in Southeast Asia. Wageningen, Alterra, Alterra-report, **2108**, ISSN 1566-7197.


