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Greenhouse gas fluxes from tropical peatlands in south-east Asia

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Abstract

The lowland peatlands of south-east Asia represent an immense reservoir of fossil carbon and are reportedly responsible for 30% of the global carbon dioxide (CO_2) emissions from Land Use, Land Use Change and Forestry. This paper provides a review and meta-analysis of available literature on greenhouse gas fluxes from tropical peat soils in south-east Asia. As in other parts of the world, water level is the main control on greenhouse gas fluxes from south-east Asian peat soils. Based on subsidence data we calculate emissions of at least 900 g CO_2 m^{-2} a^{-1} ($\sim$ 250 g C m^{-2} a^{-1}) for each 10 cm of additional drainage depth. This is a conservative estimate as the role of oxidation in subsidence and the increased bulk density of the uppermost drained peat layers are yet insufficiently quantified. The majority of published CO_2 flux measurements from south-east Asian peat soils concerns undifferentiated respiration at floor level, providing inadequate insight on the peat carbon balance. In contrast to previous assumptions, regular peat oxidation after drainage might contribute more to the regional long-term annual CO_2 emissions than peat fires. Methane fluxes are negligible at low water levels and amount to up to 3 mg CH_4 m^{-2} h^{-1} at high water levels, which is low compared with emissions from boreal and temperate peatlands. The latter emissions may be exceeded by fluxes from rice paddies on tropical peat soil, however. N_2O fluxes are erratic with extremely high values upon application of fertilizer to wet peat soils. Current data on CO_2 and CH_4 fluxes indicate that peatland rewetting in south-east Asia will lead to substantial reductions of net greenhouse gas emissions. There is, however, an urgent need for further quantitative research on carbon exchange to support the development of consistent policies for climate change mitigation.

Keywords: carbon, CH_4, CO_2, fire, greenhouse gas emissions, N_2O, peat soil, south-east Asia, subsidence, tropical peatswamp

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Introduction

The peatlands of south-east Asia, covering approximately 250 000 km^2 (Carbopeat, 2008), represent an immense reservoir of fossil carbon (Jaenicke *et al.*, 2008) that has largely accumulated over the past 13 000 years (Dommain *et al.*, 2009). Their current large scale degradation by drainage and associated peat fires has been reported to be responsible for possibly up to 30% of the global carbon dioxide (CO_2) emissions from Land Use, Land Use Change and Forestry (LULUCF; Hooijer *et al.*, 2006). The increased awareness of these CO_2 emissions has created strong political support for reducing deforestation and peatland degradation (REDD: Reducing Emissions from Deforestation and Degradation, UNFCCC, 2007), specifically in Indonesia that is responsible for the bulk of the emissions (Hooijer *et al.*, 2006). Indonesia has recently formulated a national policy with implementation regulations, in which assistance is asked for assessing greenhouse gas emissions from peatlands and for capacity building with respect to carbon accounting, baseline assessment, emission monitoring and peatland management (IFCA, 2008). The Indonesian Ministry of Forestry, with financial and technical support of Australia, Germany, the United Kingdom and the World Bank, is currently developing demonstration activities for testing and triggering a global REDD carbon market. Regrettably this political attention has not yet been paralleled by the development and acceptance of ade-

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Fig. 1 Map of the region with the approximate position of the study sites. Data on peat subsidence stem from locations 4 (Lim, 1992; Mutalib *et al*., 1992; DID & LAWOO, 1996; Wösten *et al*., 1997), 5 (Taylor & Ali, 2001), 8 (Wösten & Ritzema, 2001), 10 (CIMTROP in Hooijer *et al*., 2008), 11 (Chin & Poo, 1992; Dradjad *et al*., 2003). Data on greenhouse gas emissions stem from locations 1 (Vien *et al*., 2008), 2 (Suzuki *et al*., 1999; Ueda *et al*., 2000), 3 (Murayama & Bakar, 1996b; Ismail *et al*., 2008), 4 (Kyuma *et al*., 1992; Murayama & Bakar, 1996b), 5 (Taylor & Ali, 2001; Ali *et al*., 2006), 6 (Furukawa *et al*., 2005; Inubushi *et al*., 2005), 7 (Rumbang *et al*., 2008), 8 (Melling *et al*., 2005a, b, c, 2006, 2007a, b), 9 (Melling *et al*., 2008), 10 (Jauhiainen *et al*., 2001, 2002, 2004, 2005, 2008a, b; Darung *et al*., 2005; Takakai *et al*., 2005, 2006; Hirano *et al*., 2007a, b, 2009; Rumbang *et al*., 2008), 11 (Hadi *et al*., 2000, 2002, 2005; Inubushi *et al*., 2003, Inubushi & Hadi, 2007), 12 (Hadi *et al*., 2000, 2001, 2005), 13 (Chimner & Ewel, 2004; Chimner, 2004).

quate guidelines to assess the emissions from peatlands with the necessary accuracy.

To be tradable, under REDD or other mechanisms, including the voluntary carbon market, greenhouse gas emission reductions, e.g. from peatland rewetting and fire prevention, will have to be 'results based, demonstrable, transparent and verifiable, and estimated consistently over time' (UNFCCC, 2007). Crucial factors will be the use of rigorous and standardized methods for baseline setting and monitoring of the emissions to allow third party verification of the reductions.

This paper presents a review and meta-analysis of data related to greenhouse gas fluxes from peat soils in south-east Asia (Fig. 1) to arrive at drainage depth dependent emission rates of degraded peatlands, addressing both direct gas flux measurements as well as carbon losses from subsidence and fire events. Supplementary publications were used to gain insight on the study sites and methods. Finally, the paper identifies urgent research questions for reliably assessing the radiative forcing of natural and degraded peatlands in south-east Asia in the framework of strategies to mitigate climate change.

Results

Published information on peat subsidence rates from south-east Asian peatlands is scarce and detailed information on associated site characteristics such as water level, peat type, vegetation, land use or even location is generally lacking. Subsidence values of up to several dozen centimetres per year have been reported (Polak, 1933; Andriesse, 1988; Chin & Poo, 1992; Mutalib *et al*., 1992; DID & LAWOO, 1996); Dradjad *et al*. (2003) mention initial rates of subsidence of several centimetres per month in shallowly drained peat soils. These very high rates stem from immediately after drainage, when the peat body is compressed mechanically due to loss of supporting pore water pressure (loss of buoyancy, Schothorst, 1977; Stephens *et al*., 1984; Kennedy & Price, 2005). Leaving aside these initial high rates, Fig. 2 shows subsidence rates of tropical peat soils in relation to mean annual water level below surface. The effect of land use appears limited. Ongoing studies in *Acacia* plantations on Kampar peninsula (Sumatra; Hooijer, 2008) suggest a similar relationship between subsidence rate and water level.

We analysed the collected data on greenhouse gas fluxes from tropical peatlands of south-east Asia (CH_4, CO_2, N_2O) in their relationship to water level, pH, C/N ratio, soil temperature, vegetation cover and land use. Besides a small number of micrometeorological studies, the bulk of the published gas fluxes were measured using closed-chambers that were placed airtight on the soil. Studies differed with respect to the size and shape of chambers and the gas measurement methods (mixed headspace or not, through-flow or not), the number and frequency of gas concentration measurements made to derive fluxes and the time of day measurements were performed. An assessment of the impact of the different methods would require more controlled conditions (cf. Pumpanen *et al*., 2004; Denmead, 2008) and was not attempted.

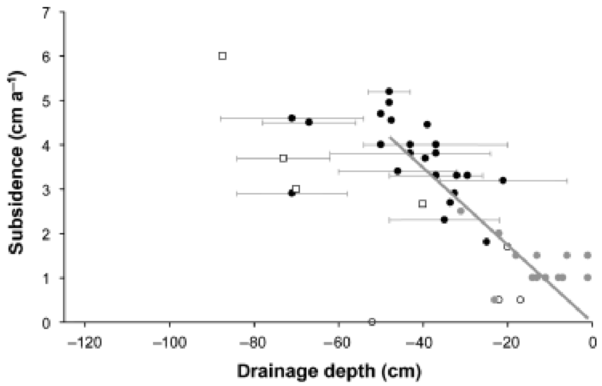

Fig. 2 Rate of subsidence in relation to mean annual water level below surface. Horizontal bars indicate standard deviation in water table (where available). Open circles denote unused, drained forested sites, these were not taken into account in the regression that applies to water levels ≤ 50 cm below surface only (slope = −0.09, $r^2 = 0.95$). Land use: (□) agriculture, from Taylor & Ali (2001), on 'old' and 'recently' cleared fields ($n = 2$), Tie & Kueh (1979), date of drainage unknown ($n = 1$) and DID & LAWOO (1996), recorded 13–18 years after drainage ($n = 1$); (●) oil palm, from DID & LAWOO (1996), recorded 13–16 or 18–21 years after drainage ($n = 23$), (●) degraded open land in the Ex Mega Rice Project area, from CIMTROP in Hooijer *et al.* (2008), recorded ~ 10–12 years after drainage ($n = 15$), (○) drained forested plots, from CIMTROP in Hooijer *et al.* (2008), recorded ~ 10–12 years after drainage ($n = 2$) and Taylor & Ali (2001), date of impact unknown ($n = 1$).

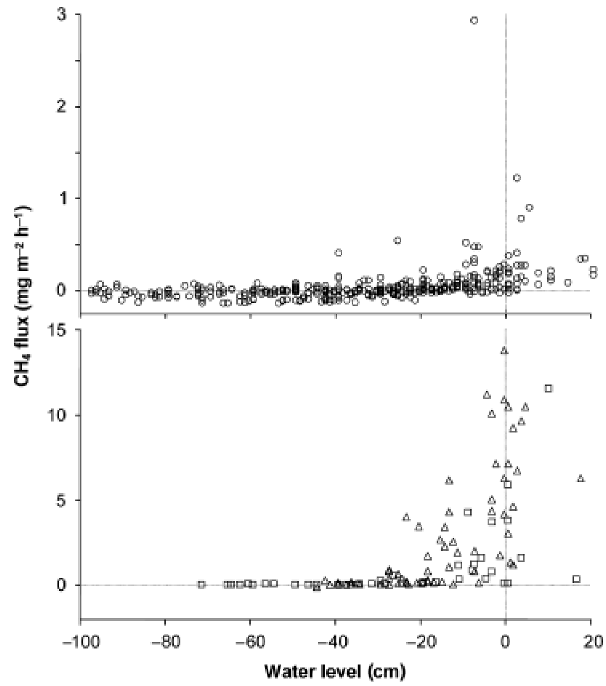

Fig. 3 Top: hourly methane fluxes from tropical peat soil in relation to water level. Negative values denote net uptake from the atmosphere by the soil. Bottom: same for (Δ) boreal and (□) temperate sites (data from: Jungkunst & Fiedler, 2007 and references therein; Augustin *et al.*, 1996; Huttunen *et al.*, 2003). Note the fivefold difference in scale.

Whereas many individual studies reveal dependencies on above mentioned site conditions, these are for the most part lost in the noise of the collected data. Nevertheless, net methane fluxes from tropical peat soils show a clear relationship to water level (Fig. 3). Values are generally low and often distinctly negative (a negative sign denotes net uptake from the atmosphere by the ecosystem) for water levels below −20 cm. At higher water levels negative values are rarer and values tend to be higher and more variable. High emissions to the atmosphere of 3.5–14 mg CH_4 m^{-1} h^{-1} are reported from paddy fields, however without indication of water levels (Hadi *et al.*, 2002, 2005). Furukawa *et al.* (2005) report emissions (mainly through the rice plant) of up to 35 mg CH_4 m^{-1} h^{-1} from paddies on peaty alluvium. In a waterlogged, previously drained and now abandoned freshwater swamp Ueda *et al.* (2000), in a series of 20 measurements with a mean of 0.5 mg CH_4 m^{-1} h^{-1}, found one aberrant value of 12.5 mg CH_4 m^{-1} h^{-1}, which they attributed to ebullition. A similar outlier of 16.8 mg CH_4 m^{-1} h^{-1} was recorded in a cassava field by Takakai *et al.* (2005). Besides high emissions from paddy fields, no differentiation between different land use types could be made. Estimates of annual net methane fluxes vary between −0.37 and 5.87 g CH_4 m^{-1} a^{-1} for forested sites, between 0.025 and 3.4 g (outlier of 12 g) CH_4 m^{-1} a^{-1} for agriculture sites and between 3.62 and 49.52 g CH_4 m^{-1} a^{-1} for rice paddies (Inubushi *et al.*, 2003; Furukawa *et al.*, 2005; Hadi *et al.*, 2005; Jauhiainen *et al.*, 2005, 2008a; Melling *et al.*, 2005a, c; Takakai *et al.*, 2005; Hirano *et al.*, 2009).

Only a handful of published micrometeorological (eddy covariance) CO_2 flux measurements are available from just five peatland sites in south-east Asia, covering a limited geographic area. Suzuki *et al.* (1999) found a net flux between August 1995 and July 1996 of ca. −1900 g CO_2 m^{-2} a^{-1} (−522 and −532 g C m^{-1} a^{-1}) in two sites in a primary and secondary peatswamp forest at To-Daeng and nearby Bacho Swamp (Narathiwat, Thailand). In near-natural but selectively logged peatswamp forest in the upper Sebangau catchment (Central Kalimantan, Indonesia), ~ 3 km from the river a net emission to the atmosphere of ~ 370 g CO_2 m^{-2} a^{-1} (~ 100 g C m^{-2} a^{-1}) was measured between May 2004 and May 2005 (Hirano *et al.*, 2007b). In a selectively logged secondary forest on drained peat in Block C of the Ex-Mega Rice Project (EMRP) area (Central Kalimantan) Hirano *et al.* (2007a, 2009) measured net emissions of 2178, 1386, 1085 and 1617 g CO_2 m^{-2} a^{-1} (594, 378, 296, 441 g C m^{-2} a^{-1}) in 2002, 2003, 2004 and 2005, respectively. Between May 2004 and May 2005, within 500 m from this selectively logged site, Hirano *et al.* (2007b) found a net emission of $\sim 2900\,g\,CO_2\,m^{-2}\,a^{-1}$ ($\sim 800\,g\,C\,m^{-2}\,a^{-1}$) in a deforested, abandoned area previously affected by fires. At ~ 1.1 km distance in the same deforested, abandoned area, Jauhiainen *et al.* (2008a) recorded similar values of $2969\,g\,CO_2\,m^{-2}\,a^{-1}$ (SD 235) for burnt, bare peat patches at mean annual water levels of -33 cm and, after rewetting, $2809\,g\,CO_2\,m^{-2}\,a^{-1}$ (SD 278) at mean annual water levels of -21 cm, using a closed dark chamber method. The vast majority of the published CO_2 flux data from south-east Asian peatlands stems from such dark chamber measurements of total (soil) respiration. These measurements cover not only heterotrophic decomposition of soil organic matter, but also autotrophic emissions from the living low vegetation as well as rhizosphere respiration. Rhizosphere respiration encompasses autotrophic activity of plant roots as well as heterotrophic activity in the rhizosphere, including decomposition of root exudates and recently dead root material (Wiant, 1967; Hanson *et al.*, 2000; Kuzyakov, 2006). Total soil respiration tends to be lower at high water tables, with forested systems showing higher values than open fields (Fig. 4).

Net nitrous oxide fluxes from soils in primary and secondary forests vary between -63 and $916\,\mu g\,N_2O\,m^{-2}\,h^{-1}$, with 90% of the measured values below $125\,\mu g\,N_2O\,m^{-2}\,h^{-1}$. With the exception of five measurements from disturbed forest sites in Central Kalimantan (Indonesia; Hadi *et al.*, 2005; Takakai *et al.*, 2006), emissions above $150\,\mu g\,N_2O\,m^{-2}\,h^{-1}$ are restricted to agricultural lands. Fluxes from fertilized agricultural sites vary between -16 and $19\,000\,\mu g\,N_2O\,m^{-2}\,h^{-1}$, with 90% of the measured values below $2000\,\mu g\,N_2O\,m^{-2}\,h^{-1}$; fluxes from abandoned sites are between -63 and $190\,\mu g\,N_2O\,m^{-2}\,h^{-1}$, with 90% below $50\,\mu g\,N_2O\,m^{-2}\,h^{-1}$. The highest emissions occur during the rainy season. Although no clear overall correlations with main site parameters (water level, C/N ratio, pH, soil temperature) could be found in the collected data, individual studies suggest that nitrous oxide fluxes are controlled mainly by land use and soil temperature and moisture conditions (Hadi *et al.*, 2000; Takakai *et al.*, 2006; Melling *et al.*, 2007b).

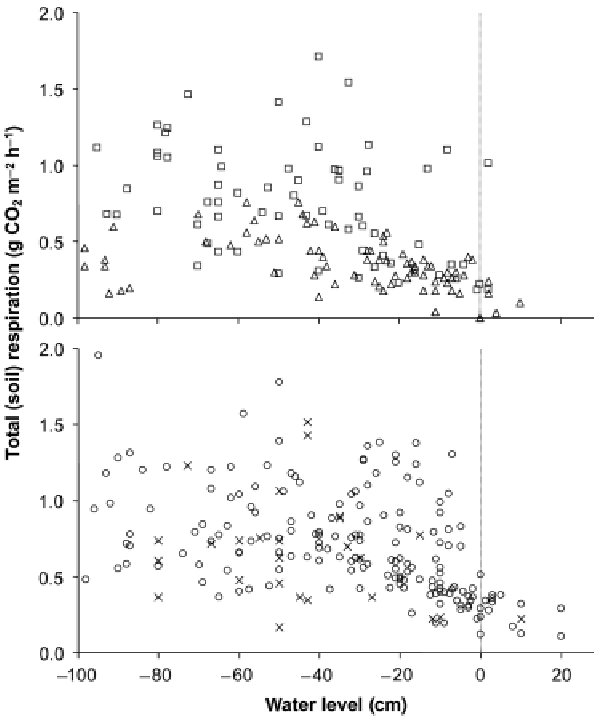

Fig. 4 Total (soil) respiration in relation to water level and land use: (□) agricultural fields; (Δ) deforested, unused; (×) agroforestry fields; (○) forested plots (includes primary, secondary, selectively logged). Note that the graphs represent total (soil) respiration not net CO_2 fluxes (see text). The large amount of soil respiration data from Hirano *et al.* (2009) and Jauhiainen *et al.* (2008a) is not included in the graphs, but falls within the same range.

Discussion

Peat subsidence is the result of several processes. In the initial stage after drainage (primary subsidence *sensu* Everett, 1983), settling or compaction occurs due to loss of supporting pore water pressure (Kennedy & Price, 2005). This initial or primary subsidence depends on type and depth of peat and the drainage level (Segeberg, 1960) and can result in drastic losses in surface height in the first years after drainage (Okey, 1918; Allison, 1946; Eggelsmann, 1960, 1978; van der Molen & Smits, 1962) as they are also found in drained tropical peatlands (Polak, 1933; Andriesse, 1988; Chin & Poo, 1992; Mutalib *et al.*, 1992; DID & LAWOO, 1996; Dradjad *et al.*, 2003). Withdrawal of water furthermore leads to shrinkage and oxidation of the now aerobic upper peat layer. In addition, wind and water erosion, leaching of soluble organic matter and fire contribute to the loss of matter and height (Eggelsmann, 1978; Everett, 1983). During the phase of secondary subsidence (Everett, 1983), shrinkage and oxidation are (in absence of fires) the dominant processes and show a linear dependency on drainage depth (Stephens & Speir, 1970; Schothorst, 1977; Eggelsmann, 1978; DID & LAWOO, 1996). When ditches are not maintained and periodically deepened to sustain desired water levels, progressive subsidence leads to increasingly thinner aerobic layers, resulting in reduced rates of subsidence (Snowden, 1986; Wösten *et al.*, 1997). Oxidation first removes easily decomposable material (Eggelsmann, 1960; Mundel, 1976; Wösten *et al.*, 1997) and also therefore oxidative losses decline with time. Tillage, fertilization and root exudates counteract this effect, resulting in continued high oxidative losses in managed agricultural peatlands (Eggelsmann, 1960; cf. Schothorst, 1977).

The observed subsidence of tropical peat soils shows the expected linear dependency on water level, at least for drainage less than 50 cm below the surface (Fig. 2). Subsidence increases by 0.9 cm a^{-1} for each 10 cm of additional drainage depth. A similar value (1.1 cm a^{-1}) is found in the Kampar study of Hooijer (2008), albeit only for water levels below −20 cm. Such a threshold water level is also present in peat subsidence studies from other parts of the world where every 10 cm of additional drainage leads to an increase in peat subsidence of 0.64 and 0.34 cm a^{-1} (Florida and Indiana, USA, respectively; Stephens & Speir, 1970), 0.43 cm a^{-1} (Balaton, Hungary; Eggelsmann, 1978) and 0.34 cm a^{-1} (Zegvelderbroek, the Netherlands; Schothorst, 1977, 1982). The limited number of observations from deeper drained tropical peatlands seems to suggest that subsidence levels off and remains at $\sim$ 4.5 cm a^{-1} at drainage depths below −50 cm (Fig. 2). In the Kampar study of Hooijer (2008), however, subsidence continues to increase up to a drainage depth of 100 cm before levelling off. A similar pattern of stabilizing (and ultimately decreasing) losses below a threshold water level is observed in annual peat oxidation rates (CO_2 emissions) of temperate European peatlands (Fig. 5). It can be ascribed to moisture stress and changes in microbial communities (Mäkiranta *et al.*, 2009).

The relative role of shrinkage in subsidence can be assessed by the increase it causes in dry bulk density of the peat (van der Molen & Smits, 1962; Schothorst, 1977, 1982; Wösten *et al.*, 1997; Dradjad *et al.*, 2003; Ewing & Vepraskas, 2006); the remainder of the total secondary subsidence is then ascribed to oxidation. As shrinkage and oxidation progressively affect increasingly deeper soil layers, depth profiles of bulk density can be used to estimate the role of shrinkage (space for time substitution). In a study on peat subsidence in western Johore (Peninsular Malaysia), DID & LAWOO (1996) used a bulk density profile of Salmah *et al.* (1992) and found that 61% of secondary subsidence could be attributed to oxidation (average of 11 calculations ranging from 54% to 73%; data from one pineapple and 10 oil palm plots); Wösten *et al.* (1997) rounded this average to 60%. These estimates fall within the (wide) range of 35–100% found in studies from other parts of the world (cf. van der Molen & Smits, 1962; Schothorst, 1977, 1982; Deverel & Rojstaczer, 1996; Schipper & McLeod, 2002; Ewing & Vepraskas, 2006; Grønlund *et al.*, 2008). If we assume that the (dark) closed chamber gas flux measurements from the burnt, bare peat patches in Block C of the Ex Mega Rice Project area (Jauhiainen *et al.*, 2008a) represent emissions from peat decomposition only (i.e. exclude rhizosphere respiration and litter decomposition) and that subsidence at this site relates to mean annual water level as deduced from Fig. 2, then, with an assumed volumetric carbon content of 0.068 g C cm^{-3} wet peat (cf. Dommain *et al.*, 2009), oxidative losses are in this case responsible for 40–60% of subsidence.

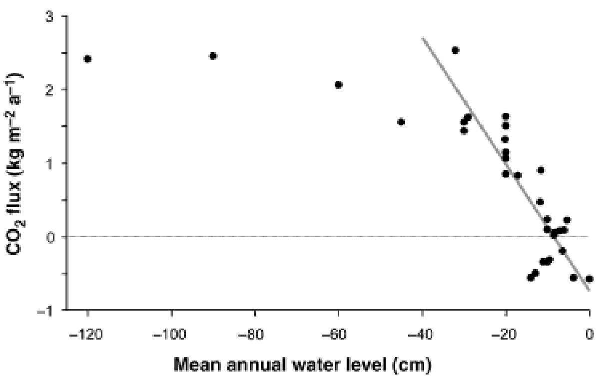

Fig. 5 Net annual CO_2 fluxes in kg CO_2 m^{-2} a^{-1} from temperate European peatlands in relation to mean annual water level. Negative values denote net uptake from the atmosphere by the ecosystem. The regression applies to water levels $\leq$ 40 cm below surface only (slope = −0.086, r^2 = 0.75). Data from Mundel (1976), Müller *et al.* (1997), Drösler (2005), Bortoluzzi *et al.* (2006), Flessa *et al.* (1997), Jacobs *et al.* (2003, 2007), Hendriks *et al.* (2007), Veenendaal *et al.* (2007).

Subsidence can be measured by remote sensing (USGS, 2003) and provides a good basis for estimating CO_2 emissions from peatland degradation if the oxidative component is known. Assuming an oxidative component of secondary subsidence of 40%, which is at the lower end of the range presented above, CO_2 emissions from drained tropical peat soils amount to 900 g CO_2 m^{-2} a^{-1} ($\sim$ 250 g C m^{-2} a^{-1}) for each 10 cm of additional drainage up to a depth of 50 cm. They may prove to be substantially larger, however. Using the 60% value, Wösten *et al.* (1997) estimate that emissions range from 1330 to 3970 g CO_2 m^{-2} a^{-1} ($\sim$ 360 to $\sim$ 1100 g C m^{-2} a^{-1}) for each 10 cm of additional drainage depth, depending on assumed bulk density of the peat. Emissions from temperate European peatlands (Fig. 5) show a similar dependency on water level of $\sim$ 900 g CO_2 m^{-2} a^{-1} for each additional 10 cm of drainage, but become positive only at water levels more than $\sim$ 10 cm below the surface and level off at a drainage depth of $\sim$ 40 cm. To arrive at more accurate and reliable estimates for tropical peatlands, more concerted subsidence studies are needed that include determination of water levels and volumetric ash and carbon content in combination with gas flux measurements.

The CO_2 data from temperate Europe (Fig. 5) represent fluxes measured over at least 12 months using reliable techniques to arrive at net fluxes between the (nonforested) ecosystem and the atmosphere. From the methodological descriptions, often quite meagre, none of the gas chamber measurements from tropical peatlands can be judged to offer data with similar reliability. With the possible exception of some of the measurements of Jauhiainen *et al.* (2008a; see above) and Melling *et al.* (2007a), none of the soil respiration studies from tropical peat soils convincingly manages to exclude rhizosphere respiration. Rhizosphere respiration is part of short-term carbon cycling and has no significant effect on carbon stocks (Kuzyakov, 2006). As rhizosphere respiration may be responsible for <10 to $>90\%$ of total soil respiration (Happell & Chanton, 1993; Hanson *et al.*, 2000; Crow & Wieder, 2005; Kuzyakov & Larionova, 2005), total soil respiration measurements are inadequate for determining the net CO_2 fluxes that contribute to global warming. Consequently, comparison of total soil CO_2 emissions with fluxes of other greenhouse gases in terms of global warming potential (GWP) (Furukawa *et al.*, 2005; Melling *et al.*, 2005a; Inubushi & Hadi, 2007; Hirano *et al.*, 2009; Jauhiainen *et al.*, 2008a) is inappropriate and should be avoided.

Soil respiration is largely confined to the upper layers of tropical peat soils (Murayama & Bakar, 1996a; Hirano *et al.*, 2009; Vien *et al.*, 2008) and, apart from autotrophic root respiration, mainly determined by decomposition of labile soil carbon (Murayama & Bakar, 1996a, b) provided by litter (Chimner & Ewel, 2005; Jauhiainen *et al.*, 2005; Yule & Gomez, 2009) and fresh root material (Chimner & Ewel, 2004). In (near-) natural tropical peatswamp forests measured soil respiration is at times higher than in degraded or agricultural sites (Melling *et al.*, 2005b; Jauhiainen *et al.*, 2008a). This is due to the higher primary production of the former (cf. Whittaker & Likens, 1973) and to pneumatophores and prop roots stimulating aerobic decomposition (Kitaya *et al.*, 2002; Chimner, 2004) of a larger amount of fresh material (Chimner & Ewel, 2004; Jauhiainen *et al.*, 2008a) by larger populations of bacteria and fungi (Hadi *et al.*, 2001). On the other hand, higher soil temperatures (Melling *et al.*, 2005b; Ali *et al.*, 2006; Ludang *et al.*, 2007) and pH (Murayama & Bakar, 1996a, b; Rumbang *et al.*, 2008) enhance decomposition in open, agricultural sites.

Micrometeorological (eddy covariance) measurements offer reliable estimates of total net ecosystem CO_2 exchange with the (local) atmosphere. Relating these to changes in soil carbon is impossible, however, without simultaneous assessment of changes in biomass and litter stocks (cf. Lohila *et al.*, 2007) and emissions of methane and dissolved and particulate carbon. This applies to disturbed and recovering forests, as well as old-growth forests (cf. Luyssaert *et al.*, 2008; Lewis *et al.*, 2009) and plantations. Closed chamber methods can be used to measure the respective contributions of surface vegetation, rhizosphere respiration, litter decomposition and peat degradation, but particularly in forested sites (includes agroforestry) these require more sophisticated set-ups (Hanson *et al.*, 2000; Kuzyakov & Larionova, 2005; Kuzyakov, 2006) than used in the available studies from tropical peatlands. Melling *et al.* (2007a) attempt to exclude rhizosphere respiration by 'trenching', i.e. inserting a cylinder into the peat severing roots well before flux measurements (Alm *et al.*, 2007; Mäkiranta *et al.*, 2008) and – without indication of drainage depth – arrive at heterotrophic soil flux rates from a 5 year old oil palm plantation of $3400–4100\,g\,CO_2\,m^{-2}\,a^{-1}$ ($930–1120\,g\,C\,m^{-2}\,a^{-1}$). The measurement method (chamber design and sampling method) used by Melling *et al.* (2007a) tends to underestimate CO_2 fluxes by 15–20% (Melling *et al.*, 2005b) or more (Norman *et al.*, 1997), however. Moreover, trenching results in increased soil moisture (Hanson *et al.*, 2000; Mäkiranta *et al.*, 2008) and reduces the stimulating 'priming effect' labile organic substances (root exudates, recent dead root material) have on decomposition of the recalcitrant peat (Mäkiranta *et al.*, 2008; cf. Hanson *et al.*, 2000; Kuzyakov *et al.*, 2000; Kuzyakov, 2006). On the other hand, trenching prevents addition of slow cycling below ground litter, but this flux seems limited in oil palm plantations (Melling *et al.*, 2007a). In conclusion, the actual flux from the 5-year-old oil palm plantation may amount to $>5000\,g\,CO_2\,m^{-2}\,a^{-1}$. In open, unforested sites, a combination of transparent and dark closed chambers and a rigorous measurement scheme allows for reliable measurements of CO_2 emissions from heterotrophic soil respiration (Drösler, 2005). Possible biases and pitfalls associated with methods chosen (chamber design, method of flux derivation, measurement frequency and time of day) and resulting uncertainties deserve more explicit attention in studies on CO_2 fluxes from tropical peatlands.

Methane emissions are restricted to high water levels when methanogenesis occurs under anaerobic conditions close to the surface and re-oxidation of methane is limited (Segers, 1998; Fig. 3). In comparison to temperate and boreal peatlands, methane emissions from tropical peatlands are low (Fig. 3). Pneumatophores in tropical peatswamps suppress methane production and stimulate methane re-oxidation by oxygen transport into the root zone. On the other hand root decay and root exudation will promote methane production (Segers, 1998). Furthermore, pneumatophores and prop roots are known to serve as methane mediators to the atmosphere in other types of forested swamps (Pulliam, 1992; Kreuzwieser *et al.*, 2003; Purvaja *et al.*, 2004) and are likely to do the same in peatswamps where also tree-aerenchyma-mediated methane emissions (Rusch & Rennenberg, 1998) should not be ruled out. But also in agricultural sites, where pneumatophores are absent, methane emissions are low, even at high water levels when re-oxidation will be limited (cf. Jauhiainen *et al*., 2001, 2004; Inubushi *et al*., 2003; Furukawa *et al*., 2005; Melling *et al*., 2005c). Under such anaerobic conditions methane may be oxidized by sulphate (Kirk, 2004), but sulphates will only be present in larger concentrations close to the often pyrite-rich mineral subsoil of tropical peatlands (Neuzil *et al*., 1993; Haraguchi *et al*., 2005) or in areas flooded by contaminated water (Miyajima & Wada, 1999; Ueda *et al*., 2000; Hadi *et al*., 2001, 2005). More likely low methane emissions from tropical peatlands relate to the poor substrate quality of the peats (high polyphenol content, e.g. lignin; Polak, 1975; Calvert *et al*., 1991; Durig & Calvert, 1991; Yonebayashi *et al*., 1992; Esterle & Ferm, 1994; Brady, 1997). Labile components are quickly depleted from the near-surface layers by aerobic decomposition (Brady, 1997; Chimner & Ewel, 2005; Yule, 2008) and lateral discharge (Yule & Gomez, 2009), and methanogenesis from the remaining more recalcitrant material is low (Miyajima *et al*., 1997; Jackson *et al*., 2008). Part of the leached organic compounds may be transported down the peat profile (cf. Waddington & Roulet, 1997), where it can serve as substrate for methanogenesis (Charman *et al*., 1994). Indeed, in bogs of north America and Europe dissolved methane and CO_2 in deeper peat layers have been found to be only about half the age of the surrounding peat matrix (Charman *et al*., 1994, 1999; Chanton *et al*., 1995, 2008; Clymo & Bryant, 2008). The young age of this methane (and CO_2) is explained by methanogenesis through reduction of CO_2 (Hornibrook *et al*., 1997; Miyajima *et al*., 1997; Chasar *et al*., 2000; Nakagawa *et al*., 2002; Clymo & Bryant, 2008; Steinmann *et al*., 2008) provided by fermentation of old peat and younger dissolved carbon (Charman *et al*., 1994; cf. Nakagawa *et al*., 2002; Chanton *et al*., 2008). As young CO_2 and CH_4 are brought in directly from upper layers as well, the dissolved gases will be of younger age than the dissolved organic compounds at the same depth (Clymo & Bryant, 2008).

The fact that the low methane emissions from tropical peatswamps are mostly derived from young sources supports the idea of limited decay of deeper peat (Chanton *et al*., 1995), as surmised from the linear age–depth relationships (cf. Dommain *et al*., 2009; cf. Gorham *et al*., 2003). Peat accumulation in the tropics is attributed to low decomposability of the material (Chimner & Ewel, 2005; Yule, 2008; Yule & Gomez, 2009) and anaerobic decomposition of old, recalcitrant material indeed may prove to be negligible.

Methane emissions from rice paddies on tropical peat soil can be substantial. Furukawa *et al*. (2005) report low emissions from peatland paddy sites at lower water levels and extremely high emissions up to $35\,mg\,CH_4\,m^{-2}\,h^{-1}$ from peaty alluvial soils at high water levels. Similarly, Hadi *et al*. (2005) found emissions of up to $14\,mg\,CH_4\,m^{-2}\,h^{-1}$ in rice paddies on peat soil. Rice provides easily degradable material as a source for the production of methane that is then transported to the atmosphere through its aerenchyma (Neue, 1993). The methane emitted from peatland rice paddies predominantly stems from acetate fermentation and is of modern age (Nakagawa *et al*., 2002). Particularly when rice straw was added, dissolved organic carbon was higher in drained tropical peatlands used for sago cultivation, resulting in increased methane production and accumulation of methane in deeper soil layers (Inubushi *et al*., 1998). Similar high concentrations of methane in the soil were observed in tropical peatlands used for sago and oil palm cultivation by Melling *et al*. (2005c), who ascribed these to higher soil temperature (Ludang *et al*., 2007) and rate of decomposition. The accumulated methane may contribute substantially to emission rates when these subsurface soil layers are opened to the atmosphere, for example through land use (Inubushi *et al*., 1998) or by ebullition (cf. Ueda *et al*., 2000). In Minnesota peatlands Glaser *et al*. (2004) observed localized and episodic large methane ebullition events when a drop in barometric pressure during periods of low water level decreased the pressure on methane pockets confined by dense wood layers. Similar processes may also take place in the woody tropical peats. Their localized extent and episodic nature make these large ebullition events hard to detect by closed chamber measurements (Glaser *et al*., 2004; cf. Comas *et al*., 2007; Denmead, 2008). Using the eddy covariance technique in a Finnish fen, Rinne *et al*. (2007) also observed highest methane fluxes during dry periods, although values were modest compared with the findings of Glaser *et al*. (2004) and Comas *et al*. (2007). Small-scale ebullition may be responsible for spikes in methane emission observed when soil moisture drops in drained tropical peat soils (cf. Furukawa *et al*., 2005; Takakai *et al*., 2005).

Whereas they display comparable erratic behaviour, nitrous oxide emissions (up to $19\,000\,\mu g\,N_2O\,m^{-2}\,h^{-1}$) and particularly consumption values (up to $-63\,\mu g N_2O\,m^{-2}\,h^{-1}$) of tropical peatland sites are large compared with values from temperate and boreal Europe (cf. Velthof *et al*., 1996; Augustin *et al*., 1998; Flessa *et al*., 1998; Münchmeyer, 2001; Maljanen *et al*., 2004; Regina *et al*., 2004; Von Arnold *et al*., 2005a, b), where peak emission values reach $\sim 5000\,\mu g\,N_2O\,m^{-2}\,h^{-1}$ (Velthof *et al*., 1996; Augustin *et al*., 1998) and peak net uptake is

Table 1 Annual nitrous oxide fluxes from peatlands in tropical south-east Asia and in temperate Europe

	Land use	$g\,N_2O\,m^{-2}\,a^{-1}$ mean (range)
Tropical south-east Asia	Drained agricultural land (fertilized), $n = 8$	14.28 (1.12–40.7)
	Drained, open vegetation (abandoned, not fertilized), $n = 5$	0.11 (−0.17–0.63)
	Forested (drained and undrained peatswamp, agro-forestry), $n = 9$	0.54 (−0.08–2.10)
	Paddy, $n = 5$	0.10 (−0.06–0.32)
Temperate Europe*	Drained agricultural land (fens/fertilized), $n = 80$	0.97 (−0.05–8.86)
	Forested (drained and undrained), $n = 14$	0.57 (0.04–2.69)
	(Semi-) natural (incl. rewetted), $n = 23$	0.10 (−0.01–0.27)

Negative values denote net uptake by the ecosystem.
*Data from Augustin & Merbach (1998); Augustin *et al.* (1998); Augustin (2003); Brumme *et al.* (1999); Drösler (2005); Flessa *et al.* (1997); Hendriks *et al.* (2007); Jacobs *et al.* (2003); Müller (1999); Tauchnitz *et al.* (2008); Velthof *et al.* (1996); Von Arnold *et al.* (2005a, b); Wild *et al.* (2001).

ca. $-8\,\mu g\,N_2O\,m^{-2}\,h^{-1}$ (Münchmeyer, 2001). Factors regulating N_2O consumption by the soil are not yet well understood and need further study (Chapuis-Lardy *et al.*, 2007). Emissions of N_2O from tropical peat soil depend on soil moisture and land use (Hadi *et al.*, 2000; Takakai *et al.*, 2006; Melling *et al.*, 2007b). The highest observed N_2O emissions were from drained and fertilized agricultural peat soils (Takakai *et al.*, 2006) and occurred when water filled pore space was between $\sim 60\%$ and $\sim 90\%$, pointing at denitrification as the main underlying process. As fertilizer was applied in form of NH_4^+-N, nitrification must also play an important role, either as direct source of N_2O, or by providing the necessary NO_3^- for denitrification (Takakai *et al.*, 2006; Hashidoko *et al.*, 2008; cf. Inubushi *et al.*, 2003; Furukawa *et al.*, 2005). Next to denitrifying bacteria (with a high potential for N_2O production; Hashidoko *et al.*, 2008), also fungi may play a major role in N_2O production in tropical peat soils (Yanai *et al.*, 2007). The role of plants in mediating N_2O emissions (cf. Rusch & Rennenberg, 1998; Kreuzwieser *et al.*, 2003) needs to be assessed. In light of the complex dependencies and resulting erratic behaviour of N_2O fluxes, measurement frequency needs to be high, particularly during the rainy season (Melling *et al.*, 2007b), to arrive at robust emission estimates. Takakai *et al.* (2006), based on year-round monthly measurements and linear interpolations, arrive at emissions from fertilized agricultural lands of 3.3–40.7 $g\,N_2O\,m^{-2}\,a^{-1}$ (with a global warming potential equivalent to 980–1210 $g\,CO_2\,m^{-2}\,a^{-1}$; cf. Forster *et al.*, 2007), which are very high compared with emissions from agricultural peatlands of temperate Europe (Table 1), by far exceed the IPPC default value of 2.5 $g\,N_2O\,m^{-2}\,a^{-1}$ (IPCC, 2006) and emphasize the need for further studies and proper land use guidelines. On the basis of the limited set of available data, primary, secondary and drained tropical peatswamp forests are indiscernible from agroforestry sites on peat with respect to N_2O emissions. Annual emissions – based on linear interpolations – from forested tropical sites are comparable to those from forested temperate European sites (Table 1).

Carbon balance

The net carbon uptake of an undrained south-east Asian primary peatswamp forest as measured using the eddy covariance technique (532 $g\,C\,m^{-2}\,a^{-1}$, Suzuki *et al.*, 1999) is an order of magnitude higher than the long-term carbon accumulation rates of tropical peatswamps as determined with palaeoecological techniques (Dommain *et al.*, 2009). This discrepancy points to additional sequestration or considerable unaccounted loss of carbon from the eddy covariance plots. A net increase in standing biomass may explain part of the difference. Export of dissolved gaseous carbon will be limited (cf. Hornibrook *et al.*, 1997; Glaser *et al.*, 2004; Clymo & Bryant, 2008; Steinmann *et al.*, 2008) and hardly affect the balance. Similarly, the amount of carbon lost through methane emissions is small. In contrast, export of dissolved organic carbon (DOC) may constitute a substantial part of the peatland carbon balance, as seen in boreal peatlands (Roulet *et al.*, 2007; Nilsson *et al.*, 2008). Tropical peatland waters can have very high DOC concentrations (Miyamoto *et al.*, 2009; Yule & Gomez, 2009) and, together with lower amounts of particulate organic carbon (POC) (cf. Yoshioka *et al.*, 2002), this carbon is exported in substantial amounts (Tachibana *et al.*, 2006; Alkhatib *et al.*, 2007; Baum *et al.*, 2007; Rixen *et al.*, 2008; cf. Hope *et al.*, 1994; Harrison *et al.*, 2005) and rapidly decomposed (cf. Hedges *et al.*, 1997; Raymond & Bauer, 2001). In order to get a clearer picture of carbon losses from tropical peatlands in relation to drainage and land use, more research on the loss of carbon through blackwater streams is

Table 2 Carbon emissions from peat fires and related parameters

Burnt peat depth (cm)	Year	Bulk density ($g\,cm^{-3}$)	Carbon content	Emission ($kg\,C\,m^{-2}$)	Fire type	Reference
37 (25–60)	1988, 1994				C	DID & LAWOO (1996)
51 (20–150)	1997	0.100*	0.57*	29.1 (11.4- 85.5)	W	Page *et al.* (2000, 2002)
55 (25–85)	1997	0.160 (0.100–0.220)	0.54 (0.53–0.56)	47.5 (13.3–104)	W	Limin *et al.* (2004)
21 (3.5–44.5)	2002	0.160 (0.100–0.220)	0.54 (0.53–0.56)	18.6 (6.3–37.1)	W	Limin *et al.* (2004)
27 (15–30)†	2002				W	Usup *et al.* (2004)
12 (0–32)	2001, 2002	0.155 (0.060–0.220)	0.50 (0.46–0.54)	9.0 (0–27.4)	E	Saharjo & Munoz (2005); Saharjo & Nurhayati (2005); Saharjo (2007)
Mean						
34		0.144	0.54	26.1		

Data are mean values (range in parentheses).
*Data from Neuzil (1997).
†Own calculations based on weight loss data.
C, clearance fire; W, wildfire; E, experimental fire.

needed. The high rate of decomposition in drained peatlands may lead to higher production of DOC that is transported out of the system through drainage canals (Holden *et al.*, 2004) together with increased POC loads from soil erosion.

In recent decades, human induced fires in south-east Asian peatlands have resulted in huge amounts of peat carbon released to the atmosphere, with single fire events resulting in losses up to well over 1 m of peat (Page *et al.*, 2000, 2002; Limin *et al.*, 2004). Based on available measurement data, the mean rate of fire-related peat loss amounts to 34 cm per fire event (Table 2; cf. Heil, 2007). This is considerably lower than the average peat loss of 51 cm measured in the EMRP area in Central Kalimantan (Indonesia) during the severe 1997/1998 El Niño drought (Table 2; Page *et al.*, 2000, 2002). Page *et al.* (2002) arrive at an average loss of 29.1 $kg\,C\,m^{-2}$ for the 1997 peatland fires (Table 2). Using a higher volumetric carbon density of 0.070 $g\,cm^{-3}$ (bulk density of 0.13 $g\,cm^{-3} \times$ carbon content of 0.54; Shimada *et al.*, 2001), Heil *et al.* (2007) arrive at a slightly higher estimate of 34.9 $kg\,C\,m^{-2}$ for the same event. Above estimates of volumetric carbon density are derived from dry bulk density and carbon content assessments over total peat depth. Because of compaction, bulk density in the upper layers of drained peatlands may be considerably higher (cf. Melling *et al.*, 2005b, 2006; Saharjo & Nurhayati, 2005; Ali *et al.*, 2006; Kool *et al.*, 2006; Kurnain *et al.*, 2006; Takakai *et al.*, 2006; Saharjo, 2007; Ywih *et al.*, 2009). Departing from a volumetric carbon content of 0.086 $g\,cm^{-3}$, Limin *et al.* (2004) report fire-related emissions in the EMRP area of 47.5 and 18.6 $kg\,C\,m^{-2}$, respectively, for the El Niño years 1997 and 2002 (Table 2). The limited number of available measurements (Table 2) implies emissions of $\sim$ 26 $kg\,C\,m^{-2}$ from a typical fire event. Compared with emissions from oxidative peat loss of 0.9 $kg\,C\,m^{-2}\,a^{-1}$ for each 10 cm of additional drainage depth, peat fire emissions are considerably larger, exceeding Holocene carbon accumulation rates by two orders of magnitude (cf. Dommain *et al.*, 2009).

While monitoring of peatland fires in south-east Asia is ongoing (Fuller & Fulk, 2001; Bechteler & Siegert, 2004; Siegert *et al.*, 2004; Hayasaka, 2007; Langner *et al.*, 2007; Miettinen, 2007; Phua *et al.*, 2007; Mastura, 2008; Putra *et al.*, 2008; Tansey *et al.*, 2008; Langner & Siegert, 2009), data on the actual volume of peat losses are scarce and adequate estimates of the relevant carbon content are lacking. Studies into both parameters are urgently needed to arrive at better estimates of carbon losses from these tropical peatland fires. If we assume a total area of $\sim 19 \times 10^9\,m^2$ of peat soil burnt during the 1997 fires in south-east Asia (Heil *et al.*, 2007), with each m^2 emitting 26 kg C (Table 2), peat carbon emissions from the 1997 fires would have amounted to $\sim$ 494 Tg C. This value corresponds well with the $\sim$ 486 Tg peat carbon emissions (67% of total emissions of 726 Tg) of van der Werf *et al.* (2008), who used CO measurements from the MOPITT satellite to optimize bottom-up estimates based on burnt area. Above values are only slightly higher than the lower estimate of 380–460 Tg peat carbon of Page *et al.* (2002). Based on van der Werf *et al.* (2008), average annual fire-related peat carbon emissions for the 2000–2006 period amount to $\sim$ 86 Tg C (67% of total emissions of 128 Tg), which is in line with the 91.5 Tg C of Heil (2007). These estimates are three to four times smaller than the lower estimate of Hooijer *et al.* (2006), who arrived at mean annual emissions from fires for the 1997–2006 period of $\sim$ 385 Tg C, with $\sim$ 340 Tg C from peat (cf. Page *et al.*,

2002) – or, recalculating for the 2000–2006 period, at $\sim$270 Tg of peat carbon.

Whereas yearly balances of oil palm plantations on tropical peat soil may suggest no or only small carbon losses (Melling *et al*., 2007a), more comprehensive longer term lifecycle analyses all arrive at clear carbon debits (Germer & Sauerborn, 2007; Pastowski *et al*., 2007; Fargione *et al*., 2008; Reijnders & Huijbregts, 2008; Wicke *et al*., 2008; Danielsen *et al*., 2009). The CO_2 emissions from peat degradation assumed in these studies range from 1.8 kg CO_2 m^{-2} a^{-1} (Germer & Sauerborn, 2007), to 3.7–5.5 kg CO_2 m^{-2} a^{-1} (Reijnders & Huijbregts, 2008), 3.9 kg CO_2 m^{-2} a^{-1} (Wicke *et al*., 2008) and 5.5–7.3 kgCO_2 m^{-2} a^{-1} (Fargione *et al*., 2008). The measurements of Melling *et al*. (2007a) indicate emission values from oxidizing peat of 5 kg CO_2 m^{-2} a^{-1} or more.

Conclusions

Subsidence in south-east Asian peatlands shows a clear linear dependency on water level, increasing by 0.9 cm a^{-1} for each 10 cm of additional drainage depth. Whereas subsidence seems to level off to remain at $\sim$ 4.5 cm a^{-1} at mean annual water levels deeper than 50 cm, recent evidence suggests there is further increase in subsidence until drainage depths of $\sim$ 1 m are attained. With a conservative estimate of 40% for the oxidative component in subsidence, CO_2 emissions amount to at least 900 g CO_2 m^{-2} a^{-1} (= 9 t CO_2 ha^{-1} a^{-1}) for each 10 cm of additional drainage depth (up to a depth of 50–100 cm). Hooijer *et al*. (2006) use a similar relationship of 910 g CO_2 m^{-2} a^{-1} and arrive at total annual emissions from peat oxidation in degrading south-east Asian peatlands of $\sim$ 600 Mt CO_2 (170 Tg C). Should emissions cease to increase with drainage depths below 50 cm, these total emissions would still amount to $\sim$ 475 Mt CO_2 (130 Tg C). More concerted studies into subsidence to arrive at better emission estimates are particularly opportune, as subsidence assessments by remote sensing may be a rapid and cheap method for monitoring CO_2 emissions from drained peatlands.

Based on the limited amount of direct measurements, 34 cm of peat is lost in an average tropical peat fire, which corresponds to 26 kg C m^{-2}. This is more than 20 times as much as the annual oxidative loss from 50 cm deep drained peat soil and exceeds average Holocene accumulation rates by 350 to over 1000 times (cf. Dommain *et al*., 2009). Applying this peat loss estimate to the total area of peat soil burnt during the 1997 fires in south-east Asia ($\sim$ 19 000 km^2; Heil *et al*., 2007), peat carbon emissions from this event would have amounted to $\sim$ 494 Tg C, which corresponds well with the $\sim$ 486 Tg of van der Werf *et al*. (2008), and is near the lower estimate of Page *et al*. (2002). Van der Werf *et al*. (2008) and Heil (2007) estimate annual fire-related peat carbon emissions for the 2000–2006 period to amount to $\sim$ 90 Tg C, which is considerably lower than the often cited estimate of Hooijer *et al*. (2006) of $\sim$ 385 Tg C. In contrast to hitherto assumed (Hooijer *et al*., 2006), regular peat oxidation after drainage on the longer run seems to contribute more to annual CO_2 emissions from south-east Asia than peat fires.

The database on CO_2 emissions from drainage and fire in tropical peatlands is still poor. Most drainage-related emission data stem from dark chamber measurements that are inadequate for determining net CO_2 fluxes to the atmosphere. Hardly any publication sufficiently considers and discusses biases, pitfalls and uncertainties associated with the flux measurement method chosen (chamber design, method of flux derivation, measurement frequency and time of day). Considering its diurnal and seasonal patterns, measurement of CO_2 fluxes must be frequent and intensive. To arrive at reliable net CO_2 fluxes between forested ecosystems and the atmosphere, rhizosphere respiration must be excluded and changes in litter and biomass stocks must be assessed. Assessing carbon stock changes furthermore involves quantifying fluxes of methane and dissolved and particulate carbon.

Total soil respiration fluxes from dark chamber measurements do not represent net-emissions and addressing them in terms of global warming potential is misleading and has led, for example, to the erroneous conclusion that oil palm plantations on drained peat are to be preferred over natural peatswamps in terms of radiative forcing (MPOC, 2006; Corley, 2007). The few passable gas measurements suggest that peat oxidation losses from drained peatlands, including oil palm plantations, reach similar values as those derived from subsidence measurements, i.e. $\sim$ 5 kg CO_2 m^{-2} a^{-1} (= 50 t CO_2 ha^{-1} a^{-1}) or more. This implies that the emission factor of biofuel derived from oil palm grown on tropical peat soil amounts to at least $\sim$ 400 g CO_2-eq MJ^{-1} (Wicke *et al*., 2008; cf. Couwenberg, 2007), which by far exceeds emission factors of common fossil fuels (cf. IPCC, 2006).

Peat losses not only lead to the emission of carbon, but also to the irrevocable loss of palaeo-archives. During the 25 year rotation period of an oil palm plantation on coastal peat soil more than 1750 years of stored information is lost. A typical fire in the EMRP area destroys at least 500 years and in extreme cases several millennia of palaeo-information. The information lost is particularly valuable as it covers the development of peatswamps in relation to Late Holocene climate variability, particularly El Niño activity, and thus provides necessary baseline information for climate mitigation projects under the REDD umbrella.

Methane emissions from south-east Asian peatlands show a clear relationship to water level. Values are generally low (and often negative) for water levels below -20 cm and higher and more variable when water levels are above this threshold. Whereas plant-mediated methane emissions need to be quantified and the possibility of periodic large scale ebullition must be assessed, in all likelihood methane emissions from tropical peatswamps are small due to the recalcitrance of the peat.

Whereas they display comparable erratic behaviour, nitrous oxide emissions from fertilized tropical agricultural peat soils are high, sometimes even extremely high, compared with those from agricultural peatlands of temperate Europe. Emissions from degraded, unfertilized, abandoned sites as well as from primary and secondary forest sites are low. More thorough measurement schemes are needed, particularly in light of the extreme emission spikes measured by Takakai *et al*. (2006).

Assessment of the global warming potential of tropical peat soils, i.e. the radiative forcing of CO_2, methane and nitrous oxide combined, as used in climate change policies (cf. Forster *et al*., 2007), depends on robust estimates of net annual emissions. With respect to land use on tropical peat soils, such robust estimates are not yet available for CO_2, the most important anthropogenic greenhouse gas (Forster *et al*., 2007), nor for N_2O.

Their predominant dependence on water level shows that rewetting of drained tropical peat soils will lead to large reductions of CO_2 emissions, also when conditions of net carbon sequestration are not reached. In contrast to temperate and boreal peatlands, the risk of substantial increase in methane emissions is low. The high N_2O emissions during the rainy season are restricted to heavily fertilized drained sites and it is highly improbable that rewetting will induce N_2O emissions that negate the CO_2 emission reductions. There is, however, an urgent need for further quantitative research on greenhouse gas exchange to support the development of consistent policies for climate change mitigation.

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