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<td>Author(s)</td>
<td>Mezbahuddin, M.; Grant, R. F.; Hirano, T.</td>
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<td>Citation</td>
<td>Biogeosciences, 11(3), 577-599</td>
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<td>2014-06-16</td>
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Modelling effects of seasonal variation in water table depth on net ecosystem CO$_2$ exchange of a tropical peatland

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Received: 3 July 2013 – Published in Biogeosciences Discuss.: 13 August 2013
Revised: 9 December 2013 – Accepted: 9 December 2013 – Published: 3 February 2014

Abstract. Seasonal variation in water table depth (WTD) determines the balance between aggradation and degradation of tropical peatlands. Longer dry seasons together with human interventions (e.g. drainage) can cause WTD drawdowns making tropical peatland C storage highly vulnerable. Better predictive capacity for effects of WTD on net CO$_2$ exchange is thus essential to guide conservation of tropical peat deposits. Mathematical modelling of basic eco-hydrological processes under site-specific conditions can provide such predictive capacity. We hereby deploy a process-based mathematical model ecosys to study effects of seasonal variation in WTD on net ecosystem productivity (NEP) of a drainage affected tropical peat swamp forest at Palangkaraya, Indonesia. Simulated NEP suggested that the peatland was a C source (NEP $\sim -2$ g C m$^{-2}$ d$^{-1}$, where a negative sign represents a C source and a positive sign a C sink) during rainy seasons with shallow WTD, C neutral or a small sink (NEP $\sim +1$ g C m$^{-2}$ d$^{-1}$) during early dry seasons with intermediate WTD and a substantial C source (NEP $\sim -4$ g C m$^{-2}$ d$^{-1}$) during late dry seasons with deep WTD from 2002 to 2005. These values were corroborated by regressions ($P < 0.0001$) of hourly modelled vs. eddy covariance (EC) net ecosystem CO$_2$ fluxes which yielded $R^2 > 0.8$, intercepts approaching 0 and slopes approaching 1. We also simulated a gradual increase in annual NEP from 2002 ($-609$ g C m$^{-2}$) to 2005 ($-373$ g C m$^{-2}$) with decreasing WTD which was attributed to declines in duration and intensity of dry seasons following the El Niño event of 2002. This increase in modelled NEP was corroborated by EC-gap filled annual NEP estimates. Our modelling hypotheses suggested that (1) poor aeration in wet soils during shallow WTD caused slow nutrient (predominantly phosphorus) mineralization and consequent slow plant nutrient uptake that suppressed gross primary productivity (GPP) and hence NEP (2) better soil aeration during intermediate WTD enhanced nutrient mineralization and hence plant nutrient uptake, GPP and NEP and (3) deep WTD suppressed NEP through a combination of reduced GPP due to plant water stress and increased ecosystem respiration ($R_e$) from enhanced deeper peat aeration. These WTD effects on NEP were modelled from basic eco-hydrological processes including microbial and root oxidation-reduction reactions driven by soil and root O$_2$ transport and uptake which in turn drove soil and plant carbon, nitrogen and phosphorus transformations within a soil-plant-atmosphere water transfer scheme driven by water potential gradients. Including these processes in ecosystem models should therefore provide an improved predictive capacity for WTD management programs intended to reduce tropical peat degradation.

1 Introduction

Seasonal and interannual fluctuations in water table depth (WTD) can affect peatland net CO$_2$ exchange through complex effects on soil oxidation-reduction reactions and hence on nutrient transformations. Shallow WTD during rainy seasons slows convective-dispersive transport of O$_2$ through wet soils. Consequent reduction in soil O$_2$ concentrations slows O$_2$ uptake used to drive aerobic oxidation-reduction reactions by soil microbes and roots and hence reduces heterotrophic and root respiration. Microbial energy yield from oxidation of reduced C coupled to reduction of O$_2$ under aerobic conditions exceeds that from oxidation coupled
to reduction of alternative electron acceptors (Brock and Madigan, 1991) under anaerobic conditions. Lower anaerobic energy yields slow microbial growth and therefore reduce heterotrophic respiration. Root oxidation-reduction reactions driving root growth and nutrient uptake also require O\textsubscript{2} which is scarce when WTD is shallow. Reduced heterotrophic and root respiration thus result in reduced ecosystem respiration ($R_e$) with shallow WTD during rainy seasons, as reported in many field studies (Cougwenberg et al., 2009; Flanagan and Syed, 2011; Limpens et al., 2008; Sulman et al., 2010). Slower microbial growth also reduces decomposition and nutrient mineralization, as well as root growth and nutrient uptake and hence gross primary productivity (GPP) (Cai et al., 2010; Flanagan and Syed, 2011; Murphy and Moore, 2010; Sulman et al., 2012).

More rapid O\textsubscript{2} transport with WTD drawdown during early dry seasons may increase root and heterotrophic respiration and hence $R_e$ (Cai et al., 2010; Sulman et al., 2010). Consequent increases in mineralization and root growth, and thereby nutrient availability and uptake can also raise GPP during this hydroperiod (Cai et al., 2010; Flanagan and Syed, 2011; Jauhiainen et al., 2012a; Sulman et al., 2012). Increased GPP in this hydroperiod may further hasten $R_e$ through increased production of fresh labile C in the forms of litter fall and root exudates (Limpens et al., 2008).

These increases, however, may not sustain with further WTD drawdown in the later part of a prolonged dry season when WTD falls below a critical depth. This critical WTD is highly site-specific depending upon peat forming vegetation and artificial drainage. For instance, this critical WTD may vary from as shallow as 0.4 m (Sonnentag et al., 2010) for pristine peatlands dominated by moss with shallow rhizoids to as deep as 0.9 m (Schwärzel et al., 2006) for drained peatlands dominated by vascular plants with deep root systems. When WTD falls below the critical depth for a particular peatland, near surface peat desiccation occurs. This desiccation can reduce near surface peat decomposition by reducing microbial access to substrate, e.g. dissolved organic C (DOC) in desiccated near-surface soil (Dimitrov et al., 2010), thereby slowing oxidation-reduction reactions and hence microbial growth. The reduction in decomposition of desiccated near surface peat can be partially or fully offset by increases in decomposition of better aerated deeper peat, thereby causing no net changes in $R_e$ during this hydroperiod (Dimitrov et al., 2010; Lafleur et al., 2005; Strack and Waddington, 2007). However, plant water stress from near surface peat desiccation might also cause a decline in GPP during deep WTD hydroperiods (Dimitrov et al., 2011; Sulman et al., 2010), thereby lowering net ecosystem productivity (NEP). Therefore, responses of peatland ecosystem net CO\textsubscript{2} exchange to WTD fluctuations are governed by basic soil hydrological and biological processes and their interactions with plant physiology.

Process-based ecosystem models can provide us with means of understanding basic mechanisms behind WTD effects on peatland net ecosystem CO\textsubscript{2} exchange. To accomplish this, a model should explicitly represent oxidation-reduction reactions, coupled with aqueous and gaseous transfers of their reactants and products. These processes require modelling WTD dynamics, soil moisture retention characteristics, gas transport through soil, differential substrate quality for microbial degradation and hydrolysis (e.g. labile vs. recalcitrant), nutrient transformations driven by these reactions, and microbial and plant nutrient uptake. However, in a review of 7 widely used ecosystem models, Sulman et al. (2012) found only ecosys (Grant, 2001) included processes to limit both CO\textsubscript{2} fixation and respiration under shallow WTD. The predictive capacity of the other models were limited by (1) not explicitly simulating WTD dynamics and consequently not modelling aerobic vs. anaerobic zones from water influxes (e.g. precipitation, lateral recharge) vs. effluxes (e.g. evapotranspiration, lateral discharge) (e.g. Kurbatova et al., 2009; St-Hilaire et al., 2010; van Huisteden et al., 2006), (2) parameterizing models with empirical rate constants and/or scalar functions for aerobic vs. anaerobic decomposition (e.g. Bond-Lamberty et al., 2007; Frolking et al., 2002; St-Hilaire et al., 2010) instead of simulating biochemical oxidation-reduction reactions affected by soil aerobicity, and (3) using scalar functions that reduce productivity in wet soils through a driver variable such as stomatal conductance ($g_s$) (e.g. Bond-Lamberty et al., 2007; Frolking et al., 2002) instead of simulating nutrient limitations to CO\textsubscript{2} fixation imposed by reduced nutrient availability and root nutrient uptake caused by slower oxidation-reduction reactions resultant of slower O\textsubscript{2} transport processes through soils and roots. The general purpose terrestrial ecosystem model ecosys includes site-independent algorithms representing all the processes affected by aerobicity mentioned above, thereby excluding the need for arbitrary model parameterization. The model could therefore successfully simulate WTD effects on $R_e$ and GPP of different peatlands without site-specific parameterization (e.g. Dimitrov et al., 2010, 2011; Grant et al., 2012a).

All of the previous peatland modelling studies mentioned above have been tested only against measurements from northern temperate and boreal peatlands. Modelling the fate of vulnerable C storage in tropical peatlands under WTD fluctuations is still largely under-investigated. For instance, modelling WTD effects on tropical peat soil respiration has to date predominantly included regressions of soil CO\textsubscript{2} fluxes against WTD (e.g. Couwenberg et al., 2009; Hirano et al., 2009; Hooijer et al., 2010; Jauhiainen et al., 2008, 2012b; Melling et al., 2005) without taking other confounding factors like land use, nutrient availability, nature of the peat, ecosystem productivity etc. into consideration (Murdiyarso et al., 2010). Modelling eco-physiological response to hydrology in tropical peatlands is particularly important since climates in tropical peatlands are very different from those in northern temperate and boreal peatlands. Tropical peatlands are formed under high temperature and precipitation,
an important consequence of which is that a small WTD drawdown might cause a large increase in peat decomposition (Page et al., 2009). Distinct dry seasons almost every year together with human intervention such as drainage have been reported to deepen WTD, thereby causing rapid decomposition of very old and thick (up to > 26 000 yr old and 9 m thick as C dated and measured by Page et al., 2004) tropical peat deposits (Couwenberg et al., 2009; Hirano et al., 2009, 2012; Jauhiainen et al., 2008). Moreover, tropical peatlands are generally formed by roots and remains of trees and devoid of bryophytes (e.g. mosses) as opposed to the northern peatlands that are mainly formed by mosses or co-dominated by mosses and vascular plants. Trees have well developed root systems and stomatal regulations that are lacking in bryophytes. These differences can alter plant water and nutrient uptake processes in tree dominated peatlands from those in bryophyte dominated peatlands. Consequently WTD effects on productivity of tropical peatlands may be very different than of those in northern boreal and temperate peatlands. Besides, tropical peat deposits formed by tree remains can have very different substrate quality for microbial decomposition than boreal and temperate moss peatlands and hence may have a different WTD – peat respiration interaction. This difference in peat forming materials can also cause different hydrological characteristics thereby producing very different water retention and transport phenomena between tropical and temperate/boreal peatlands. Variations in climate and peat forming vegetation thus necessitate rigorous testing of process models against measurements across peatlands developed under very different climate (e.g. boreal vs. tropical) and vegetation (e.g. moss vs. tree) to improve predictive capacity for eco-hydrological controls on peatland C balance.

The process-based hourly time step ecosystem model *ecosys* previously simulated the effects of WTD fluctuations on net CO$_2$ exchange of northern boreal peatlands (Dimitrov et al., 2011; Grant et al., 2012a). Testing the same model against site measurements of a tropical peatland would thus be an important test of the versatility of its algorithms representing the processes described above. Such a test will allow us to determine whether our current understanding of peatland water, nutrient and C interactions is sufficiently robust to capture complex WTD effects on peatland $R_c$ and GPP over a wide range of climates (boreal to tropical). Our study hereby uses *ecosys* to simulate WTD effects on net CO$_2$ exchange of a drainage affected tropical peat swamp forest at Palangkaraya, Central Kalimantan, Indonesia (Hirano et al., 2007). These effects are summarized in modelling hypotheses during three seasonal hydroperiods as follows:

1. Shallow WTD in the rainy season (November–April) causes lower net ecosystem productivity (NEP) mainly through slower CO$_2$ fixation due to reduced nutrient availability and uptake caused by slower nutrient trans-

formation and root growth and uptake resulting from slower O$_2$ diffusion through wet soils.

2. When WTD increases during the early dry season (May–July), more rapid O$_2$ transport into larger unsaturated soil zones enables faster root growth and microbial nutrient transformations that in turns results in more rapid root nutrient uptake and CO$_2$ fixation which contributes to a higher NEP. Increased O$_2$ availability in this hydroperiod may, however, result in more rapid aerobic decomposition in deeper peat layers. Drying of surface residues and near surface peat layers at the same time can reduce surface and near surface soil respiration thereby offsetting the increase in deeper peat respiration, resulting in no net increase of $R_c$.

3. Deeper WTD during the late dry season (August–October), causes greater desiccation of near surface peat which forces declines in root and canopy water potentials, and consequently in canopy conductance and CO$_2$ fixation, thereby reducing NEP. Further deepening of the aerobic peat zone during this hydroperiod may lead to an increase in deeper peat respiration which exceeds reduction in near surface peat respiration through desiccation, raising $R_c$ and further lowering NEP.

2 Methods

2.1 Model development

2.1.1 General

*Ecosys* is a general purpose terrestrial ecosystem model that simulates 3-D soil-microbes-plant-atmosphere water, energy, C and nutrient (nitrogen, phosphorus) transfer schemes (Grant, 2001). Details of the key algorithms in *ecosys* are given in the supplementary materials from Appendices A to H with definitions of the terms and references. Algorithms governing WTD effects on net ecosystem CO$_2$ exchange that are directly related to our modelling hypotheses are described in the following sections. For simplicity and clarity we have not put the equations within the text. Instead we have cited the equations within brackets with the letter representing a particular appendix in the supplementary materials.

2.1.2 Heterotrophic respiration

Organic transformations in *ecosys* occur in five organic matter-microbe complexes (course woody litter, fine non-woody litter, animal manure, particulate organic C (POC), and humus), each of which consists of five organic states (three decomposition substrates: solid organic C, sorbed organic C and microbial residue C, as well as the decomposition product: DOC, and the decomposition agent: microbial
biomass) in a surface residue layer and in each soil layer. The decomposition rates of each of the three substrates and resulting production of DOC in each complex is a first-order function of the active biomasses (M) of diverse heterotrophic microbial functional types, including obligate aerobes (bacteria and fungi), facultative anaerobes (denitrifiers), obligate anaerobes (fermenters), heterotrophic (acetotrophic) and autotrophic (hydrogenotrophic) methanogens, and aerobic and anaerobic heterotrophic diazotrophs (non-symbiotic N2 fixers) (A1, A2). Decomposition rates are calculated from the fraction of substrate mass colonized by M(A4). Growth of M by each microbial functional type (A25) is calculated from its uptake of DOC (A21), driven by energy yields from growth respiration (Rg) (A20) remaining after subtracting maintenance respiration (Rm) (A18) from heterotrophic respiration (Rh) (A11) driven by DOC oxidation (A13). This oxidation may be limited by microbial O2 reduction (A14) driven from microbial O2 demand (A16) and constrained by O2 diffusion calculated from aqueous O2 concentrations in soil ([O2g]) (A17). Values of [O2g] are maintained by convective-dispersive transport of O2 from the atmosphere to gaseous and aqueous phases of the soil surface layer (D15), by convective-dispersive transport of O2 through gaseous and aqueous phases in adjacent soil layers (D16, D19), and by dissolution of O2 from gaseous to aqueous phases within each soil layer (D14a).

With shallower WTD during the rainy season, air-filled porosity (θg) above the water table may decline to values at which low O2 diffusivity in the gaseous phase (Dg) (D17) may reduce gaseous O2 transport (D16), while θg below the water table is zero and so prevents gaseous O2 transport. During this hydroperiod, [O2g] relies more on O2 transport through the slower aqueous phase (D19). A consequent decline in [O2g] slows O2 uptake (A17) and hence Rh (A14), Rg (A20) and growth of M (A25). Lower M in turn slows decomposition of organic C (A1, A2) and production of DOC which further slows Rg (A21), and hence M growth, organic C decomposition and subsequent DOC production. Slower decomposition of organic C under low [O2g] also causes slower decomposition of organic nitrogen and phosphorus (A7) and production of dissolved organic nitrogen (DON) and phosphorus (DOP), which causes slower uptake of microbial nitrogen and phosphorus (A22) and hence growth of M (A29). This slower growth causes slower mineralization of organic nitrogen and phosphorus, and hence lower aqueous concentrations of NH4+, NO3− and H2PO4− (A26).

Increase in θg with WTD drawdown during the dry season results in greater Dg (D17) and hence more rapid gaseous O2 transport. A consequent rise in [O2g] increases O2 uptake (A17) and hence Rh (A14), Rg (A20) and growth of M (A25). Larger M in turn hastens decomposition of organic C (A1, A2) and production of DON which further hastens Rg (A13), Rh and growth of M. More rapid decomposition of organic C under adequate [O2g] in this hydroperiod also causes more rapid decomposition of organic nitrogen and phosphorus (A7) and production of DON and DOP, which increases uptake of microbial nitrogen and phosphorus (A22) and hence growth of M (A29). This rapid growth causes rapid mineralization of organic nitrogen and phosphorus, and hence greater aqueous concentrations of NH4+, NO3− and H2PO4− (A26).

However, desiccation of surface litter and near surface soil resulting from deepening WTD decreases litter and soil water contents and potentials (ψa) which cause an increase in aqueous microbial concentrations ([M]) (A15). This reduces microbial access to the substrate for decomposition through an algorithm for competitive inhibition of microbial exoenzymes (A4) from Lizana and Suzuki (1990), thereby reducing Rh (A13).

2.1.3 Autotrophic respiration and growth

Growth of root and shoot phytomass in each plant population is calculated from its assimilation of the non-structural C product of CO2 fixation (σc) (C20). Assimilation is driven by Rh (C17) remaining after subtracting Rm (C16) from autotrophic respiration (Ra) (C13) driven by oxidation of σc (C14). This oxidation in roots may be limited by root O2 reduction (C14b) which is driven by root O2 demand to sustain C oxidation and nutrient uptake (C14e), and constrained by O2 uptake controlled by concentrations of aqueous O2 in the soil ([O2g]) and roots ([O2r]) (C14d). Values of [O2g] are maintained by convective-dispersive transport of O2 through soil gaseous and aqueous phases and by dissolution of O2 from soil gaseous to aqueous phases. Values of [O2g] are maintained by convective-dispersive transport of O2 through the root gaseous phase (D16d) and by dissolution of O2 from root gaseous to aqueous phases (D14b) through processes analogous to those described under Sect. 2.1.2. This transport depends on species-specific values used for root air-filled porosity (θpr) (D17b).

Low θg with shallow WTD during the rainy season reduces soil O2 transport, forces root O2 uptake to rely more on [O2g] and hence on root O2 transport determined by θpr. If this transport is inadequate, decline in [O2g] slows root O2 uptake (C14c, d) and hence Ra (C14b), Rg (C17) and root growth (C20b). Increased θg with WTD drawdown during the dry season, however, facilitates rapid Dg which allows root O2 demand to be almost entirely met from [O2g] (C14c, d).
2.1.4 Gross primary productivity

By reducing root O2 uptake, shallow WTD slows root growth (C20b) and root nitrogen and phosphorus uptake (C23b, d, f). Root nitrogen and phosphorus uptake in this hydroperiod is further slowed by reductions in aqueous concentrations of NH4+, NO3− and H2PO4− (C23a, c, e) from slower mineralization of organic nitrogen and phosphorus as described in Sect. 2.1.2. Slower root nitrogen and phosphorus uptake in turn reduces concentrations of non-structural nitrogen and phosphorus products of root uptake (σN and σP) with respect to that of σC in leaves (C11), thereby slowing CO2 fixation (C6) and hence GPP.

Increased availability of [O2] with WTD drawdown during the dry season hastens root O2 uptake and so enables more rapid root growth and nitrogen and phosphorus uptake as discussed in Sect. 2.1.3. Increased root growth and nitrogen and phosphorus uptake is further stimulated by increased aqueous concentrations of NH4+, NO3− and H2PO4− (C23a, c, e) from more rapid mineralization of organic nitrogen and phosphorus during this hydroperiod as described in Sect. 2.1.2. Greater root nitrogen and phosphorus uptake in turn increases concentrations of σN and σP with respect to σC in leaves (C11), thereby facilitating rapid CO2 fixation (C6) and hence GPP.

With deeper WTD during the late dry season, GPP is less limited by root and microbial growth and nutrient uptake as discussed above. However, GPP in this hydroperiod can be adversely affected by water stress. When WTD deepens past a critical depth (Sect. 1), inadequate capillary rise (D9a) causes near-surface peat desiccation, reducing soil water potential (ψs) and increasing soil hydraulic resistance (Ωs) (B9), forcing lower root, canopy and turgor potentials (ψr, ψc and ψt) (B4) and hence lower gs (B2b) to be calculated when equilibrating plant water uptake with transpiration (T) (B14). Lower gs in turn reduces CO2 diffusion into the leaves thereby reducing CO2 fixation (C6) and hence GPP during this hydroperiod.

Thus WTD effects on Rh, Ra and GPP in ecotone are not parameterized from ecosystem level observations, but instead are governed by basic processes of O2 transport and uptake; root and microbial energy yields, growth and nutrient uptake; and stomatal regulation controlled by root water uptake parameterized from independent research.

2.2 Modelling experiment

2.2.1 Site conditions

The ecotone algorithms for simulating WTD effects on net ecosystem CO2 exchange were tested against the measurements over a drainage affected tropical peat swamp forest at Palangkaraya, Central Kalimantan, Indonesia (2°20′42″ S and 114°2′11″ E). The site is a tropical ombrotrophic bog peatland formed mainly by roots and remains of trees where the major source of water and nutrient inputs is through precipitation. Vegetation of these peatlands includes evergreen over-storey trees and dense under-storey of dominant tree seedlings with no mosses. These peatlands were drained by excavating drainage canals during 1996–1997, approximately 5 yr before the measurements started at a flux station established in Palangkaraya Drained Peat Swamp Forest (PDPSF). Peat depth around the flux tower site was about 4 m. Detail description of edaphic and vegetation characteristics as well as management history of the site can be found in Hirano et al. (2007) and Jauhiainen et al. (2008).

2.2.2 Field datasets

CO2 fluxes used for model validation in our study were measured by Hirano et al. (2007) at a flux station established in PDPSF during November 2001. Hourly NEP (a negative sign represents an upward flux or a flux out of the ecosystem and a positive sign represents a downward flux or a flux into the ecosystem) over PDPSF was estimated by a combination of eddy and storage CO2 flux measured using a micrometeorological approach (Hirano et al., 2007). NEP along with latent heat (LE) and sensible heat (H) fluxes were measured using an open path CO2/H2O analyzer mounted at 41.3 m height, about 15 m above the forest canopy (Hirano et al., 2007). A CO2 profile was also measured by using a closed path analyzer at six heights between 2.0 and 41.3 m. Hourly weather variables (e.g. incoming longwave and shortwave radiation, wind speed, relative humidity, air temperature, precipitation etc.) were also measured at the flux station. Soil moisture content (θ) at a depth from 0–0.2 m and soil temperature (Ts) at 0.05 m depth were measured in hummocks. Hourly WTD measurements at the site were started from April 2004. Ground reference point for WTD measurements was a hollow surface.

A u* (friction velocity) threshold of 0.17 m s−1 was used to screen out the NEP in a calm night hour. However, no u* threshold screening was performed for daytime NEP. NEP measured in the rain was also excluded (Hirano et al., 2005, 2012). Nighttime NEP that survived the quality control was used as a measure of nighttime Re. Daytime Re was extrapolated by using nighttime Re. GPP was then calculated by adding daytime Re to daytime NEP that survived the quality screening.

Flux gaps due to quality control were filled by Hirano et al. (2007) through look up tables (LUT) created for four periods of 3 months each (November–January, February–April, May–July and August–October). Hourly measured θ, WTD and Ts were used to incorporate environmental controls in LUTs for filling nighttime NEP (= Re) gaps. No hourly WTD measurement was available at the site before April 2004. Therefore, only θ was used as hydrological control in LUTs from January 2002–March 2004. In order to balance the number of original data in each cell, θ was grouped into six classes from below 0.23 m3 m−3 to above 0.35 m3 m−3
with an interval of 0.03 m$^3$m$^{-3}$. $T_h$ was also grouped into five classes from 25 to 30 °C with an interval of 1 °C. In filling GPP gaps, photosynthetic photon flux density (PPFD) and vapour pressure deficit ($D$) were used as environmental factors to create LUTs similar to those used for $R_e$. PPFD and $D$ were grouped into eight classes from 5–250 to above 1750 µmol m$^{-2}$ s$^{-1}$ at intervals of 250 µmol m$^{-2}$ s$^{-1}$ and three classes of below 1.2, 1.2–1.8 and above 1.8 kPa, respectively. Finally, NEP gaps during daytime hours were calculated by the differences between gap-filled GPP and estimated daytime $R_e$.

A more detailed description of eddy covariance (EC) methodology, measurement techniques, quality control, and partitioning and gap filling of fluxes can be found in Hirano et al. (2007, 2012).

2.2.3 Model run

For our modelling experiment, the PDPSF landscape was represented by one hummock and one hollow grid cell each of which had a dimension of 1 m x 1 m. Both of the grid cells had identical soil properties except that the hollow grid cell had a fibric layer thinner by 0.15 m than the hummock cell to represent the average site micro-topography described by Jauhiainen et al. (2008) (Table 1). Physical and hydrological characteristics of the PDPSF peatland and their representation in our modelling study were described in our concurrent paper on modelling seasonal variation in WTD and surface energy exchange over PDPSF (Mezbahuddin et al., 2014). Very high carbon and nitrogen to phosphorus ratios (C : N and C : P) with low pH are typical characteristics of tropical peatlands which were represented in our modelling experiment by inputs measured either at the same site or at similar surrounding sites (Table 1). Both hummock and hollow grid cells were seeded with evergreen tropical rainforest over- and under-storey vascular vegetation using the same plant functional types used in an earlier study on an Amazonian rainforest (Grant et al., 2009), but selecting 0.2 for root porosity ($\theta_r$) used in root O$_2$ transport (D17d) to represent wetland plant adaptation (Visser et al., 2000). These peatlands under study are generally devoid of mosses and hence we did not simulate any moss species. The model was then run for 44 yr (40 yr of spin up and 4 yr of simulation run) under repeating 4 yr sequences of hourly weather data (solar radiation, air temperature, wind speed, humidity and precipitation) recorded at the site from 2002 to 2005. The spin up period allowed CO$_2$ exchange in the model to achieve stable values through successive weather sequences. Model results for the 4 yr of simulation run were compared with measurements at PDPSF from 2002–2005.

2.2.4 Model validation

Hourly CO$_2$ fluxes modelled over the hummock and the hollow were spatially averaged to represent 50 : 50 hummock-

hollow ratios as described by Jauhiainen et al. (2008), and then regressed on hourly measured EC CO$_2$ fluxes for each year from 2002–2005. Model performance was evaluated from regression intercepts ($a \rightarrow 0$), slopes ($b \rightarrow 1$) and coefficients of determination ($R^2 \rightarrow 1$) for each study year in order to test whether there was any systematic divergence between the modelled and EC measured as well as between modelled and gap-filled CO$_2$ fluxes. This test is very important since any small divergence between hourly modelled and EC measured as well as between hourly modelled and gap-filled CO$_2$ fluxes can result in a large divergence between modelled and EC-gap filled annual estimates.

2.2.5 Analyses of model results

In order to examine WTD effects on seasonal variations in NEP as proposed in hypotheses 1, 2 and 3 of our study, we chose daily modelled and EC-gap filled NEP for 30 days each from the three WTD hydroperiods, i.e. shallow, intermediate and deep, for 2002–2005. These 30 day periods were chosen based on the greatest availability of EC measured CO$_2$ fluxes that passed the quality control procedure described in Sect. 2.2.2. We then performed single factor analyses of variance (ANOVA) for the modelled and EC-gap filled NEP to test whether the means of daily NEP significantly differed among different hydroperiods. A significant difference in mean NEP between two particular hydroperiods meant the variation in mean NEP between those hydroperiods was larger than the day to day variation in NEP within each of those hydroperiods. This test would signify the consistency of seasonal variations in NEP as a result of WTD fluctuations.

2.2.6 Model sensitivity to drained vs. undrained WTD

WTD in the modelled grid cells in ecosys arises from water exchanges with the atmosphere in the forms of vertical water influxes (e.g. precipitation) and effluxes (e.g. evapotranspiration) through a surface boundary and in the forms of recharge and discharge with an adjacent ecosystem through lateral boundaries. The distance ($L_t$) and hydraulic gradient between modelled WTD and a set external water table depth (WTD$_s$, representing mean WTD of the adjacent watershed) generally governs the rate of lateral recharge and discharge (D10, D10a). All modelled WTDs were spatially averaged for the hummock and the hollow grid cell with reference to the hollow surface. The WTD$_s$ for the simulation in this study was set at 0.45 m below the hollow surface (i.e. 0.60 m below the hummock surface) so as to represent the average watershed WTD measured for our drained site (Hirano et al., 2012). $L_t$ was set to 400 m in all directions which was the nearest distance from the study site to the drainage canal (Hirano et al., 2012). Since drainage is a key disturbance reported to alter WTD and hence C balance of Southeast Asian peatlands (Couwenberg et al., 2009; Hooijer et
Table 1. Key soil properties as *ecosys* inputs to represent a drainage affected tropical peat swamp forest at Palangkaraya, Indonesia *.

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<th>$D_{\text{humm}}$ (m)</th>
<th>$D_{\text{holl}}$ (m)</th>
<th>TOC (g kg$^{-1}$)</th>
<th>TON (g Mg$^{-1}$)</th>
<th>TP (g Mg$^{-1}$)</th>
<th>pH</th>
<th>CEC (cmol + kg$^{-1}$)</th>
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<tr>
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<td>2.00</td>
<td>49</td>
<td>3.75</td>
<td>37.5</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* $D_{\text{humm}}$ = depth from hummock surface, $D_{\text{holl}}$ = depth from hollow surface, TOC = total organic C (maximum limit of input for TOC concentration in *ecosys* is used from an average of TOC values for top 4 m of a 9 m deep tropical peat column measured by Page et al., 2004), TON = total organic nitrogen and TP = total phosphorus (values obtained from Page et al., 1999), pH and CEC = cation exchange capacity (values obtained from Sayok et al., 2007).

3 Results

3.1 Modelled vs. measured ecosystem net CO$_2$ fluxes

Regressions of hourly modelled vs. measured net ecosystem CO$_2$ fluxes gave intercepts within 1.0 µmol m$^{-2}$ s$^{-1}$ of zero, and slopes within 0.1 of one, indicating minimal bias in modelled values for all years of the study except 2005 when modelled fluxes gave a positive bias slightly greater than 1.0 µmol m$^{-2}$ s$^{-1}$ (Table 2). Values for coefficients of determination ($R^2$) and root-mean-square for errors (RMSE) were $\sim$0.8 ($P < 0.0001$) and $\sim$ 5.0 µmol m$^{-2}$ s$^{-1}$ (Table 2). Much of the unexplained variance in EC-measured CO$_2$ fluxes could be attributed to a random error of ca. 20% in EC methodology (Wesely and Hart, 1985). This attribution was corroborated by root-mean-squares for random errors (RMSRE) in EC measurements, calculated for forests with similar CO$_2$ fluxes from Richardson et al. (2006) that were similar to RMSE. These similar values indicated that further constraint in model testing could not be achieved without further precision in EC measurements. Regressions of modelled vs. gap-filled CO$_2$ fluxes gave larger slopes than those of modelled vs. EC-measured CO$_2$ fluxes despite higher $R^2$ and lower RMSEs, indicating the diurnal variation of the modelled CO$_2$ fluxes was systematically larger than that of the gap-filled CO$_2$ fluxes (Table 2). Further investigation into hourly simulated vs. gap-filled and simulated vs. EC measured net ecosystem CO$_2$ fluxes suggested that modelled nighttime fluxes were systematically larger than the gap-filled nighttime fluxes particularly in the rainy season when valid EC measured data were scarce (Fig. 1). However, modelled nighttime fluxes showed good agreement with more available valid EC measured nighttime fluxes during the dry season (Fig. 1).

3.2 Seasonal variation in WTD and daily net ecosystem CO$_2$ exchange

WTD in PDPSF showed distinct seasonality in each year from 2002 to 2005 (Mezbahuddin et al., 2014). Observed WTDs were typically within 0.3 m of the hollow surface during the rainy season (November–April) increasing to...
0.5–0.8 m below the hollow surface at the onset of the dry season (May–July) (Figs. 2–5). During late dry seasons (August–October) observed WTD fell below 1.0 m from the hollow surface (Figs. 2–5). Increasing amounts and declining seasonality of precipitation caused the wet to dry season drawdown of WTD to be more and more gradual from the driest year 2002 to the wettest year 2005 (Figs. 2–5).

NEP (a negative sign represents C source and a positive sign represents C sink) modelled and measured over PDPSF showed a distinct seasonality, with negative values over shallow WTD (within 0.3 m below the hollow surface) during the rainy season, near zero or slightly positive values over intermediate WTD (0.5–0.8 m below the hollow surface) during the early dry season, and returning to negative values over deep WTD (>1.0 m below the hollow surface) in the late dry season during each year from 2002 to 2005 (Figs. 2–5). These values indicated that the ecosystem was a C source when the WTD was shallow, became C neutral or a small sink when WTD receded to an intermediate position, and again became a large source of C when WTD further deepened (Figs. 2–5).

Modelled and EC-gap filled NEP during intermediate WTD hydroperiods were significantly ($P < 0.01$) higher than those during shallow WTD hydroperiods during 2002–2003 (Fig. 6). Both the modelled and EC-gap filled NEP showed similar trend of increasing NEP from shallow to intermediate WTD hydroperiods during 2004–2005. However, only the increases in modelled NEP from shallow to intermediate WTD hydroperiods in those years were statistically significant ($P < 0.01$) (Fig. 6). Both modelled (in 2002–2004) and EC-gap filled (in 2002–2005) NEP during deep WTD hydroperiods was significantly ($P < 0.01$) lower than that during intermediate WTD hydroperiods (Fig. 6). This seasonal trend in NEP also varied interannually depending upon the duration and intensity of the dry seasons. For instance, NEP during the deep WTD hydroperiod was more negative in a drier dry season (2002–2004, Figs. 2–4 and 6) than that in a wetter dry season (2005, Figs. 5–6).

In addition to the successful simulations of interannual variation in seasonal cycles of NEP, ecosys was adequately sensitive to the short-term variations in NEP caused by changes in weather. There were several short-term dips in EC-gap filled NEP, e.g. DOY 160–170 in 2002 (Fig. 2), DOY 258–262 in 2003 (Fig. 3), DOY 143–146 in 2004 (Fig. 4), DOY 259–262 in 2005 (Fig. 5) etc. caused by smaller CO2 influxes and larger CO2 effluxes on cloudy and rainy days. These dips were modelled from less CO2 fixation under lower $R_n$ and/or from flushes of soil CO2 effluxes due to rewetting surface residues from a rainfall following a dry period (Grant et al., 2012b).

### 3.3 Seasonal variation in WTD and diurnal CO2 exchange

In order to examine WTD effects on ecosystem diurnal net CO2 exchange, we compared hourly modelled net CO2 fluxes against EC-gap filled CO2 fluxes binned for the three WTD hydroperiods, i.e. shallow, intermediate and deep from 2002–2005 (Fig. 7). During 2002–2003, modelled downward CO2 fluxes were suppressed over shallow WTD during the rainy seasons, became larger over intermediate WTD during the early dry seasons, and again suppressed over deeper WTD during the late dry seasons, as were also apparent in EC-gap filled CO2 fluxes (Fig. 7). During 2004, modelled downward fluxes followed the same seasonal pattern as in 2002–2003 but EC-gap filled fluxes showed clear suppression in only the deep WTD hydroperiod (Fig. 7). Both modelled and EC-gap filled downward CO2 fluxes during 2005, however, increased with deepening WTD with no suppression during late dry season (Fig. 7). Suppressions of downward CO2 fluxes during deep WTD periods varied interannually depending upon the duration and intensity of the hydroperiods. For instance, suppression of modelled and EC-gap filled downward CO2 fluxes during the deep WTD hydroperiods was stronger during the drier late dry seasons of 2002 and 2004, less strong in wetter late dry season of 2003 and absent in the wettest late dry season of 2005 (Fig. 7). Limited precision and frequency of EC-measured nighttime CO2 fluxes caused by insufficient nighttime turbulence made the comparison of modelled vs. EC-gap filled upward CO2 fluxes more difficult than that for downward CO2 fluxes. EC-gap filled upward CO2 fluxes ($= R_e$) showed no significant change except a small decrease during 2003 with WTD drawdown from rainy to early dry seasons and marked increases with further deepening of WTD in late dry seasons during 2002–2005 (Fig. 7). These effects of seasonal WTD variation on upward CO2 fluxes ($= R_e$) were reasonably well simulated except that the modelled decrease in upward CO2 fluxes from shallow to intermediate WTD hydroperiod in 2003 was smaller than that in EC-gap filled fluxes (Fig. 7). Moreover, modelled upward CO2 fluxes ($= R_e$) had a small decrease from rainy to dry season in 2005 which was not apparent in EC-gap filled fluxes (Fig. 7). Increase in EC-gap filled upward CO2 fluxes ($= R_e$) from intermediate to deep WTD hydroperiod was not prominent in modelled upward fluxes during 2002 (Fig. 7).

Both modelled and EC-gap filled downward CO2 fluxes increased from shallow to intermediate WTD hydroperiods during 2002, 2003 and 2005 while upward CO2 fluxes ($= R_e$) decreased little or not at all, suggesting that GPP was raised by gradual drawdown of WTD from the rainy to early dry season (Fig. 7). An increase in $R_n$ from the rainy to early dry season may have contributed to the increase in downward CO2 fluxes during 2002 (Figs. 2b and 7). However, a similar increase in $R_n$ from the rainy to early dry seasons during 2003 and 2004 did not coincide with a similar
Table 2. Modelled vs. measured net ecosystem CO$_2$ fluxes over a drainage affected tropical peat swamp forest at Palangkaraya, Indonesia$^a$.

<table>
<thead>
<tr>
<th>Year</th>
<th>Precipitation (mm yr$^{-1}$)</th>
<th>n</th>
<th>a</th>
<th>b</th>
<th>$R^2$</th>
<th>RMSE</th>
<th>RMSRE</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Modelled vs. EC measured CO$_2$ fluxes recorded at $u^* &gt; 0.17$ m s$^{-1}$</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2002</td>
<td>1852</td>
<td>3007</td>
<td>0.82</td>
<td>1.03</td>
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<td>5.5</td>
</tr>
<tr>
<td>2003</td>
<td>2291</td>
<td>2595</td>
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<td>1.05</td>
<td>0.83</td>
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<td>5.9</td>
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<td>2004</td>
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<td>3299</td>
<td>0.61</td>
<td>1.01</td>
<td>0.83</td>
<td>4.9</td>
<td>5.8</td>
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<tr>
<td>2005</td>
<td>2620</td>
<td>3164</td>
<td>1.09</td>
<td>1.01</td>
<td>0.81</td>
<td>5.2</td>
<td>5.6</td>
</tr>
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<td></td>
<td>Modelled vs. gap-filled CO$_2$ fluxes</td>
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<td></td>
<td></td>
<td></td>
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<td></td>
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<tr>
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<tr>
<td>2003</td>
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<td>−0.35</td>
<td>1.10</td>
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<td></td>
</tr>
</tbody>
</table>

$^a$ $(a, b)$ from simple linear regressions of modelled on measured. $R^2 =$ coefficient of determination and RMSE = root-mean-square for errors from simple linear regressions of measured on modelled. All measured values were recorded at $u^* > 0.17$ m s$^{-1}$. RMSRE = root-mean-square for random errors in EC measurements calculated by inputting EC fluxes recorded at $u^* > 0.17$ m s$^{-1}$ into algorithms for estimation of random errors in EC measurements developed for forests by Richardson et al. (2006).

Fig. 1. Hourly EC measured, gap-filled and simulated net ecosystem CO$_2$ fluxes and observed and simulated soil water contents ($\theta$) from 0–0.20 m depth of the hummock during 2002 and 2003 over a drainage affected tropical peat swamp forest at Palangkaraya, Indonesia.

increase in downward flux, whereas no change in $R_n$ between those hydroperiods during 2005 coincided with an increase in downward CO$_2$ fluxes (Figs. 3–5 and 7). Moreover, an increase in $D$ from rainy to early dry season (Fig. 6f) did not cause a decline in downward CO$_2$ fluxes during 2002–2005 (Fig. 7). These confounding effects of $R_n$ and $D$ on
modelled and EC-gap filled downward CO₂ fluxes further indicated that there was a consistent increase in GPP from shallow to intermediate WTD hydroperiods which was driven by eco-hydrology rather than micrometeorology. Larger declines in downward modelled and EC-gap filled CO₂ fluxes than the increases in upward CO₂ fluxes (= Rₑ) from intermediate to deep WTD hydroperiods during 2002–2004 indicated GPP suppression by deep WTD (Fig. 7). Higher D (e.g. during 2002–2004 in Fig. 6f) and lower Rₑ (e.g. due to smoke haze shading from surrounding forest and peat fires during 2002 in Fig. 2b as mentioned by Hirano et al., 2007) may have further contributed to GPP suppression during deep WTD hydroperiods (Fig. 7). However, interannual variation in the intensity of GPP suppression (Fig. 7) coincided with that in the duration and intensity of dry seasons irrespective of changes in D and Rₑ (Figs. 2–6), as described earlier, further suggested the significance of hydrological control over micrometeorological control in suppressing GPP during deep WTD hydroperiods.

Therefore, a gradual drawdown of WTD from rainy to early dry season resulted in higher NEP mainly by raising GPP with no change or little decrease in Rₑ (Figs. 2–7). Further drawdown of WTD during late dry season forced NEP to decline by a combination of reduced GPP and increased Rₑ (Figs. 2–7). This seasonal effect of WTD on NEP through its effects on GPP and Rₑ can also be corroborated by the quadratic curve fittings between monthly modelled and EC-derived NEP, GPP and Rₑ vs. monthly modelled and observed WTD. Such curve fittings between NEP and WTD yielded goodness of fits (R²) of 0.61 (modelled NEP vs. modelled WTD) and 0.53 (EC-derived NEP vs. observed WTD) indicating a small increase in NEP from shallow to intermediate WTD hydroperiods and a remarkable decline in NEP from intermediate to deep WTD hydroperiods during 2002–2004 (Fig. 8). The quadratic relationships between GPP and WTD (R² = 0.14 for modelled GPP vs. modelled WTD and R² = 0.10 for EC-derived GPP vs. observed WTD) indicated that increases in GPP from shallow to intermediate WTD hydroperiods contributed to increases in NEP and declines in GPP from intermediate to deep WTD hydroperiods contributed to declines in NEP (Fig. 8). Rₑ and WTD relationships (R² = 0.53 for modelled Rₑ vs. modelled WTD and R² = 0.60 for EC-derived Rₑ vs. observed WTD) indicated that no change or small decreases in Rₑ from shallow to intermediate WTD hydroperiods contributed little to increases in NEP, and large increases in Rₑ from intermediate to deep WTD hydroperiods contributed substantially to declines in NEP (Fig. 8). Though monthly modelled
NEP values were similar to EC-gap filled NEP values, both monthly modelled GPP and $R_e$ were systematically larger than EC-derived GPP and $R_e$ (Fig. 8).

### 3.4 Interannual variation in WTD and NEP

Interannual variation in WTD over PDPSF from 2002 to 2005 was mainly caused by differences in annual precipitation. Increasing amount of annual precipitation from the driest year 2002 to the wettest year 2005 drove a gradually shallower average annual modelled and measured WTD from 2002 to 2005 (Table 3). Variation in annual estimates of neither modelled nor EC-gap filled evapotranspiration (ET) from 2002 to 2005 did correlate with that in average annual modelled or observed WTD (Table 3). However, modelled lateral discharge increased from 2002 to 2005 with decreasing WTD (Table 3). This interannual variation in WTD caused interannual variation in NEP over PDPSF from 2002 to 2005. A gradual rise in both modelled and EC-gap filled annual NEP was found from 2002 to 2004 with progressively shallower WTD (Table 3). However, modelled NEP was considerably lower than the EC-gap filled estimates of NEP in 2003 and 2004 (Table 3). The decreasing WTD from 2002 to 2005 reduced modelled annual GPP, $R_a$ and $R_h$ and hence $R_e$ (Table 3), although this reduction could not be corroborated from EC-derived estimates. In contrast, EC-derived annual GPP increased with decreasing WTD from 2002–2004 but EC-derived $R_e$ showed no response (Table 3). Moreover, modelled annual GPP and $R_e$ were consistently larger than the EC-derived estimates during 2002–2004 (Table 3). Similar declines in modelled GPP and $R_a$ with shallower WTD left modelled NPP almost unchanged throughout the study period (Table 3). However, greater suppression of annual $R_h$ by shallow WTD caused annual NEP to become gradually less negative from 2002 to 2005 (Table 3).

### 3.5 Seasonal and annual variations in simulated drained vs. undrained WTD and NEP

Large negative simulated and EC-gap filled annual NEP during 2002–2005 (Table 3) may reflect disturbance effects of drainage in 1996–1997 which increased WTD. In order to examine the drainage effects on modelled NEP, we performed a drained vs. undrained model sensitivity test as described in Sect. 2.2.6. During the rainy seasons (November–April) from 2002–2005, simulated undrained WTD was always above the hollow surface as opposed to the simulated drained WTD. In contrast, during the dry seasons (May–October) from 2002–2005, simulated undrained WTD was always below the hollow surface as opposed to the simulated drained WTD.
WTD where water table never rose above the hollow surface (Fig. 9). The undrained WTD remained \( \sim 0.5 \text{ m} \) shallower than the drained WTD, and so altered the timing and intensity of the different hydroperiods (Fig. 9). The seasonal variation in simulated undrained WTD followed that in a nearby similar undrained tropical peat swamp forest (Fig. 9).

NEP modelled in the undrained condition was higher (less negative) than that in the drained condition during the rainy seasons (November–April) but similar during the late dry seasons (August–October) (Fig. 10). Large spikes of negative NEP were simulated in the undrained simulation at the end of the rainy seasons when WTD first declined below the hollow surface (Fig. 10). This decline suddenly increased contact between atmosphere and aqueous CO\(_2\) in the previously saturated soil, causing rapid degassing. Such spikes were not found in the drained simulation where WTD remained below the hollow surface.

On an annual basis, the undrained simulation produced a shallower average WTD by \( \sim 0.5 \text{ m} \) than the drained simulation over four years i.e. 2002–2005. This reduction in WTD in undrained vs. drained simulation decreased GPP and \( R_a \) slightly, but decreased \( R_h \) much more, thereby increasing mean annual NEP throughout the study period (Table 4).

### Table 3. Simulated (sim) and observed (obs) annual water and C balance over a drainage affected tropical peat swamp forest at Palangkaraya, Indonesia \(^a\).

<table>
<thead>
<tr>
<th>Units</th>
<th>Precipitation</th>
<th>ET</th>
<th>( Q )</th>
<th>Avg. WTD</th>
<th>GPP</th>
<th>( R_a )</th>
<th>NPP</th>
<th>( R_h )</th>
<th>( R_e )</th>
<th>NEP</th>
</tr>
</thead>
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<tr>
<td></td>
<td>mm yr(^{-1})</td>
<td></td>
<td></td>
<td>m</td>
<td>g C</td>
<td>m(^{-2}) yr(^{-1})</td>
<td>yr(^{-1})</td>
<td>yr(^{-1})</td>
<td>g C yr(^{-1})</td>
<td>yr(^{-1})</td>
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<tr>
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<td>obs</td>
<td>sim</td>
<td>obs</td>
<td>obs</td>
<td>sim</td>
<td>obs</td>
<td>other studies</td>
</tr>
</tbody>
</table>

\(^a\) ET = evapotranspiration, observed ET for each year was calculated from EC-gap filled hourly latent heat fluxes measured by Hirano et al. (2005); \( Q \) = total lateral discharge; WTD = water table depth, simulated and observed WTDs are averages of data used in Figs. 2c, 3c, 4c and 5c for 2002, 2003, 2004 and 2005 respectively; observed mean WTD for 2002 was not calculated due to the lack of site measurements and/or literature values for the first six months (Fig. 2c); GPP = gross primary productivity, observed GPP values were derived by Hirano et al. (2007) from EC-gap filled net CO\(_2\) flux partitioning (Sect. 2.2.2); \( R_a \) = autotrophic respiration; NPP = net primary productivity (NPP = GPP – \( R_a \)); \( R_h \) = heterotrophic respiration; \( R_e \) = ecosystem respiration, observed \( R_a \) values were derived by Hirano et al. (2007) from EC-gap filled net CO\(_2\) flux partitioning (Sect. 2.2.2); and NEP = net ecosystem productivity (NEP = NPP – \( R_a \)), observed NEP values were derived from hourly EC-gap filled net ecosystem CO\(_2\) fluxes measured and gap-filled by Hirano et al. (2007) (Sect. 2.2.2).

\(^b\) for Amazonian forest (Chambers et al., 2004).

\(^c\) for oil palm plantations in tropical peatlands of Malaysia (Melling et al., 2008).

\(^d\) for a mature Acacia plantation in a drained Indonesian peatland with an average WTD of 0.8 m (Jauhiainen et al., 2012b).

### 4 Discussion

#### 4.1 Modelling hypotheses of WTD effects on seasonal variation in tropical peatland NEP

Reduction of NEP during both shallow and deep WTD hydroperiods with respect to that in the intermediate WTD hydroperiods was established during 2002–2005 in Sects. 3.2 and 3.3. Reduction of NEP during the shallow WTD hydroperiods was mainly attributed to reduction in GPP that was independent of changes in \( R_a \) and \( D \) (Sect. 3.3). Reduction of NEP during the deep WTD hydroperiods, however, was attributed to reduction in GPP irrespective of changes in \( R_a \) and \( D \), and to increase in \( R_h \), irrespective of changes in temperature (variations in mean daily air temperature and mean daily soil temperature measured at 0.05 m depth of the hummocks were less than 3 °C among the hydroperiods) (Sect. 3.3). The absence of a decline in GPP during the deep WTD hydroperiod in the wettest year 2005 also suggested that there was a considerable interannual variation in WTD effect on tropical peatland NEP that depended on the intensity and duration of dry vs. wet seasons (Sect. 3.3). Seasonal variation in WTD thus affected that in NEP through its effect on both GPP and \( R_h \) as mentioned above independent of variations in other micrometeorological controls such as \( R_a \), \( D \) and temperature. These effects suggested that tropical peatland NEP was reduced by plant processes as affected by soil hydroperiods.
processes influenced by both shallow and deep WTD. Since our modelling could reasonably simulate seasonal variation in WTD as well as its effects on that in NEP (Sects. 3.2 and 3.3), we hereby discuss the modelling hypotheses proposed at the beginning of our study to explain the underlying causes of these WTD effects on tropical peatland NEP.

Table 4. Sensitivity of modelled annual C balance to drainage over a drainage affected tropical peat swamp forest at Palangkaraya, Indonesia*.

<table>
<thead>
<tr>
<th>Units</th>
<th>Year</th>
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<th>undr</th>
<th>dr</th>
<th>undr</th>
<th>dr</th>
<th>undr</th>
<th>dr</th>
<th>undr</th>
<th>dr</th>
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<tbody>
<tr>
<td>WTD m</td>
<td>2002</td>
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<td>0.22</td>
<td>0.73</td>
<td>0.17</td>
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<td>−0.10</td>
<td>0.59</td>
<td>0.10</td>
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<td>GPP g C m−2</td>
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<td>4040</td>
<td>3665</td>
<td>4123</td>
<td>3680</td>
</tr>
<tr>
<td>Rₐ g C yr⁻¹</td>
<td></td>
<td>2909</td>
<td>2412</td>
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<td>2455</td>
<td>2760</td>
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<td>2396</td>
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<td>1292</td>
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<td>1325</td>
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<td>−118</td>
<td>−373</td>
<td>−51</td>
<td>−497</td>
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</tbody>
</table>

*dr = drained simulation; undr = undrained simulation (Sect. 2.2.6); WTD = water table depth, values are the annual means of simulated WTD data in Fig. 9, positive values represent WTDs below hollow surface and the negative values represent WTD above the hollow surface; GPP = gross primary productivity; Rₐ = autotrophic respiration; NPP = net primary productivity (NPP = GPP − Rₐ); Rₑ = heterotrophic respiration; Rₑ = ecosystem respiration; and NEP = net ecosystem productivity (NEP = NPP − Rₑ).

Fig. 6. (a) Average simulated daily net ecosystem productivity (NEP sim) (b) average EC-gap filled daily net ecosystem productivity (NEP EC-gap) (c) average simulated water table depths (WTD sim) (d) average observed water table depths (WTD obs) (e) total observed precipitation (P), and (f) average mid-day (10:00–14:00 LT) vapour pressure deficit (D) during shallow (DOY 1–30, 41–70, 11–40 and 41–70 in 2002, 2003, 2004 and 2005 respectively), intermediate (DOY 121–150, 146–175, 181–210 and 121–150 in 2002, 2003, 2004 and 2005 respectively) and deep (DOY 251–280, 251–280, 291–320 and 241–270 in 2002, 2003, 2004 and 2005 respectively) water table depth (WTD) hydroperiods (delineated by vertical dotted lines in Figs. 2–5) over a drainage affected tropical peat swamp forest at Palangkaraya, Indonesia. Bars represent standard errors of means. Asterisks (***) at the top of a column represent significant (P < 0.01) difference from the adjacent column(s).

Fig. 7. Hourly binned simulated (lines) and EC-gap filled (symbols) ecosystem net CO₂ fluxes during shallow (DOY 1–30, 41–70, 11–40 and 41–70 in 2002, 2003, 2004 and 2005 respectively), intermediate (DOY 121–150, 146–175, 181–210 and 121–150 in 2002, 2003, 2004 and 2005 respectively) and deep (DOY 251–280, 251–280, 291–320 and 241–270 in 2002, 2003, 2004 and 2005 respectively) water table depth (WTD) hydroperiods (delineated by vertical dotted lines in Figs. 2–5) over a drainage affected tropical peat swamp forest at Palangkaraya, Indonesia. Solid symbols are averages of 30 hourly values more than 1/3 of which were EC measured CO₂ fluxes recorded at u* (friction velocity) > 0.17 m s⁻¹. Open symbols are averages of 30 hourly values less than 1/3 of which were EC measured CO₂ fluxes recorded at u* > 0.17 m s⁻¹. Bars represent standard errors of means.
Hollow surface depths below the hollow surface and positive values mean depths 2003, 2004 and 2005 respectively. Negative values of WTD mean WTDs are the same as those in Figs. 2c, 3c, 4c and 5c for 2002, tropical peat swamp forest at Palangkaraya, Indonesia. Drained (WTD) (Sect. 2.2.6) during 2002–2005 over a drainage affected Biogeosciences, 11, 577–

Fig. 9. Simulated daily drained vs. undrained water table depths (WTD) (Sect. 2.2.6) during 2002–2005 over a drainage affected tropical peat swamp forest at Palangkaraya, Indonesia. Drained WTDs are the same as those in Figs. 2c, 3c, 4c and 5c for 2002, 2003, 2004 and 2005 respectively. Negative values of WTD mean depths below the hollow surface and positive values mean depths above the hollow surface.

4.1.1 Hypothesis 1: WTD and NEP during rainy season

Fig. 10. Five-day moving averages of simulated daily net ecosystem productivity (NEP) for drained vs. undrained simulations (Sect. 2.2.6) during 2002–2005 over a drainage affected tropical peat swamp forest at Palangkaraya, Indonesia. Drained values are the same as those in Figs. 2–5.

A shallower aerobic zone (Fig. 11) and resulting lower \([O_2]_r\) modelled during the shallow WTD (within 0.3 m of the hollow surface) hydroperiod reduced rates of C oxidation by microbial populations (A13–A14) which limited microbial growth (A25) and hindered nutrient mineralization (A26). Moreover, low \([O_2]_r\) caused by shallow WTD forced \([O_2]_r\) to depend predominantly upon \(O_2\) transport through root gaseous phase controlled by \(\theta_{pr}\) as discussed in Sect. 2.1.3. Our input of 0.20 for \(\theta_{pr}\) to represent peatland species adaptation (Visser et al., 2000), however, was not enough to maintain adequate \([O_2]_r\) below the water table during this hydroperiod. Lower \([O_2]_r\) suppressed rates of C oxidation by root and mycorrhizal populations (C14a, b), thereby slowing root and mycorrhizal growth (C20b) below the water table. Consequently, root and mycorrhizal growth during this hydroperiod were largely confined to the shallow aerobic zone, thus limiting the soil volume from which nutrient (predominantly phosphorus) uptake (C23b, d, f) could occur. Slow phosphorus uptake reduced \(\sigma_P\) with respect to \(\sigma_C\) in leaves (C11), thereby slowing CO\(_2\) fixation (C6) and hence GPP in the model as discussed in Sect. 2.1.4. The suppression of productivity with shallow WTD was also apparent in higher modelled and measured Bowen ratios \((\beta = H/LE)\) resulting from lower \(g_s\), required to conserve the ratio of CO\(_2\) concentration at the photosynthetic site to that in the atmosphere \((C_o : C_a)\) with slower CO\(_2\) diffusion (Mezbahuddin et al., 2014). Reduction of GPP in our undrained simulation compared to our drained simulation was also caused by greater phosphorus limitation under shallower WTD (Table 4).

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In tropical peatlands, phosphorus is likely to be more limiting to CO₂ fixation than nitrogen as indicated by foliar nitrogen to phosphorus ratios (N : P) ranging from 22 : 1 to 130 : 1 measured for different tree species growing in Indonesian peatlands (Tuah et al., 2000). A mass based foliar nitrogen to phosphorus ratio greater than 16 : 1 is thought to indicate phosphorus limitation to plant productivity over nitrogen limitation (Aerts and Chapin, 2000). This phosphorus limitation arises from low plant phosphorus availability resultant of high soil organic nitrogen to phosphorus ratios (N : P), low pH and high aluminum (Al) and iron (Fe) contents of these peats (Page et al., 1999, 2006; Rieley and Page, 2005). When site-specific inputs for soil pH, organic and inorganic nitrogen and phosphorus (Table 1), exchangeable Al (12 g Mg⁻¹) and Fe (21.6 g Mg⁻¹) were used in our modelling with site-independent algorithms for soil solute transformations (E1-E55) (Grant et al., 2009), soil solution phosphorus concentrations and hence root phosphorus uptake were forced to very low values. Low plant phosphorus availability and uptake in the model was reflected in lower modelled foliar phosphorus content (∼2.5 g kg⁻¹ C) during shallow WTD hydroperiods. Such low phosphorus status in our modelling was corroborated by even lower foliar phosphorus contents (0.2–1.4 g kg⁻¹ C in matured leaves, assuming that 50% of the dry matter was C) measured by Tuah et al. (2000) in the dominant tree species growing on our study site. Low foliar phosphorus contents in 150 tree species of tropical rain forests of Costa Rica and Brazil was also attributed to low phosphorus availability for plants growing on highly weathered phosphorus-deficient tropical soils (Townsend et al., 2007).

These results from the model are also consistent with those from Milner (2009) from a transect study on the effect of soil fertility on vegetation diversity of a tropical mixed swamp forest surrounding our study site. She found a reduction in basal area and tree growth in areas where the WTD was shallower by only ∼0.10 m. She speculated that the reduced tree growth was a result of low nutrient availability caused by anoxia under shallow WTD. Slow nutrient (phosphorus) mineralization in wet soils under shallow WTD may have reduced CO₂ diffusion and thereby fixation and productivity in her study site which supports the nutrient stress theory in our hypothesis 1. Moreover, a reduction in foliar phosphorus content during the rainy season with respect to mid-wet and dry seasons was also measured by Townsend et al. (2007) in tropical forest species of Costa Rica. Furthermore, tropical mangroves in Panama, Belize and Florida have shown significant increases in leaf CO₂ assimilation and plant growth with phosphorus enrichment indicating phosphorus stress to leaf gas exchange under anoxic conditions despite those species being well adapted to flooding stress (Lovelock et al., 2004, 2006a, b and c). Similar suppression of productivity caused by low nitrogen availability and uptake can also be found in boreal peatlands which are known to be more nitrogen than phosphorus limited as reported in experimental (Flanagan and Syed, 2011; Sulman et al., 2010) and modelling studies (Grant et al., 2012a; Sulman et al., 2012) across northern boreal peatlands.

During shallow WTD in rainy seasons, [O₂s] below the water table was almost zero (Fig. 11) and well below the Michaelis-Menten constant (Kₘ) used for microbial, root and mycorrhizal uptake in ecosys (A17a, C14c) so that O₂ diffusion becomes constrained by the low [O₂s]. It is expected that DOC oxidation coupled with O₂ reduction was strongly limited by [O₂s]. In these layers, DOC oxidation was coupled with DOC reduction by anaerobic heterotrophic fermenters, which yielded much less energy (4.4 vs. 37.5 kJ g C⁻¹) (A21) than did O₂ reduction and so resulted in slower microbial growth (A25) and Rₘ (A13) as discussed in Sect. 2.1.2. However, [O₂s] above the water table during shallow WTD hydroperiod was well above Kₘ (Fig. 11) used for simulated microbial, root and mycorrhizal O₂ uptake so that Rₘ in this zone was not limited by [O₂s]. Moreover, frequent precipitation throughout this hydroperiod kept the surface residue layer moist and maintained optimum heterotrophic microbial concentrations for decomposition (A3, A5) and growth (A15), driving surface CO₂ flushes. This absence of suppression in modelled Rₑ during shallow WTD hydroperiods can further be corroborated by no significant decrease or even slight increase in nighttime EC-gap filled CO₂ fluxes during shallow compared to intermediate WTD hydroperiods (Fig. 7). Similar absence of suppression in soil respiration during shallow WTD hydroperiods compared to those in intermediate WTD hydroperiods were also measured by
Sundari et al. (2012) in our study site and Jauhiainen et al. (2008) in a nearby drained tropical peatland.

Shallow WTD during the rainy season thus caused lower NEP (Figs. 2–6) to be modelled over PDPSSF through reducing GPP due to plant nutrient (phosphorus) stress and hence net CO$_2$ uptake (Figs. 7–8). This GPP suppression from nutrient stress during shallow WTD hydroperiod was well corroborated by EC-gap filled CO$_2$ fluxes (Fig. 7) (Table 2) as well as other biometric measurements (e.g. Townsend et al., 2007) and hence validated our hypothesis 1.

4.1.2 Hypothesis 2: WTD and NEP during early dry seasons

Increased availability of [O$_2$] with an intermediate WTD (0.5–0.8 m below the hollow surface) during the early dry season almost entirely met root O$_2$ demand and hence facilitated more rapid and deeper root growth (C20b) and hence phosphorus uptake (C23b, d, f). Uptake was further stimulated by more rapid mineralization of organic phosphorus (C23a, c, e) driven by more rapid microbial O$_2$ uptake (A17), C oxidation (A1, A2), growth (A25) and $R_h$ (A13, A20) as described in Sects. 2.1.2 and 2.1.3. Greater root phosphorus uptake in turn increased $\sigma_P$,$\sigma_C$ in leaves (C11), thereby facilitating rapid CO$_2$ fixation (C6) and hence GPP as discussed in Sect. 2.1.4. Rapid CO$_2$ fixation from improved plant phosphorus status was also apparent in larger EC-gap filled downward net CO$_2$ fluxes during intermediate WTD hydroperiods in 2002, 2003 and 2005 (Fig. 7). Rapid CO$_2$ fixation due to improved plant phosphorus status can further be corroborated by lower measured and modelled $\beta$ resulting from greater $g_s$ as described in our concurrent paper on modelling WTD effects on surface energy exchange over PDPSSF (Mezbahuddin et al., 2014).

Foliar phosphorus content was reported as a good indicator of soil phosphorus availability in many studies (e.g. Aerts and Chapin, 2000; Townsend et al., 2007). Modelled foliar phosphorus content increased from $\sim 2.5$ g kg$^{-1}$ C in the wet season to $\sim 3$ g kg$^{-1}$ C in the early dry season indicating increased soil phosphorus availability and subsequent increased plant phosphorus uptake in the early dry season with intermediate WTD. This increase in modelled foliar phosphorus content by $\sim 20\%$ from rainy to early dry season was consistent with a $25\%$ increase in foliar phosphorus content from wet to mid-wet and dry seasons measured by Townsend et al. (2007) in tropical forest species of Costa Rica. Similarly, higher foliar phosphorus content modelled in the drained simulation ($\sim 3$ g kg$^{-1}$ C) than in the undrained simulation ($\sim 2.5$ g kg$^{-1}$ C) during this hydroperiod raised GPP, further indicating improved phosphorus status due to deeper WTD (Table 4). Increased productivity resulting from improved nutrient (phosphorus) availability with WTD drawdown has been found in field studies for Indonesian peatlands (Milner, 2009), and Florida everglades wetlands (Saha et al., 2010). These field measurements further support our modelling hypothesis 2 of improved plant nutrient (phosphorus) status with improved soil aeration which increased GPP and hence NEP during intermediate WTD hydroperiods. Increased GPP with WTD drawdown was also measured by Sulman et al. (2009) and Flanagan and Syed (2011) for northern boreal peatlands. However, in those ecosystems, the drawdown of WTD could improve plant nitrogen rather than phosphorus availability and uptake and hence productivity as modelled with $ecosys$ by Grant et al. (2012).

A deeper aerobic zone and resulting increase in [O$_2$] during the intermediate WTD hydroperiod in the early dry season (Fig. 11) stimulated $R_h$ (A13, A20) as described above. However, this increase in deeper $R_h$ was fully, and sometimes more than fully, offset by decreases in surface and near-surface $R_h$ caused by near-surface peat desiccation which reduced microbial access to substrate for decomposition (A15) (Sect. 2.1.2). This enabled $ecosys$ to simulate nighttime net CO$_2$ fluxes measured by EC during intermediate WTD hydroperiods that were similar to or lower than those in shallow WTD hydroperiods. This modelling hypothesis in $ecosys$ was further corroborated by soil CO$_2$ effluxes measured with surface chambers by Sundari et al. (2012) over our study site and by Jauhiainen et al. (2008) over a nearby similar site during intermediate WTD hydroperiods that were similar to or lower than those in shallow WTD hydroperiods. Declines in $R_h$ due to near surface peat desiccation were also modelled by Dimitrov et al. (2010) and Grant et al. (2012a) by using the same model $ecosys$ over two contrasting northern boreal peatlands.

Intermediate WTD during the early dry season thus caused higher NEP (Figs. 2–6) over PDPSSF by a combination of increased GPP and unchanged or slightly decreased $R_e$ (Figs. 7–8). This trend of increased GPP with improved nutrient (phosphorus) status and unchanged $R_e$ due to offsetting effects of surface vs. deep respiration was well corroborated by EC-gap filled CO$_2$ fluxes (Fig. 7) (Table 2) and other field measurements in our study site and in similar ecosystems and hence validated our hypothesis 2.

4.1.3 Hypothesis 3: WTD and NEP during late dry seasons

GPP during the late dry season with deep WTD (> 1.0 m below the hollow surface) was limited not by plant nutrient status but by plant water stress. Inadequate recharge of near surface peat layers through a combination of less precipitation and slow capillary rise (D9a) during this hydroperiod reduced $\psi_s$ and increased $\Omega_s$ (B9) in those layers where most of the plant roots were located (Mezbahuddin et al., 2014). These changes in turn forced higher $\Omega_t$ (B10), lower $\psi_r$, $\psi_c$, $\psi_t$ (B4) and $g_s$ (B2b) (Mezbahuddin et al., 2014) and hence slower CO$_2$ diffusion (C6) through stomata and consequently less GPP (Sect. 2.1.4). GPP suppression due to plant water stress can further be corroborated by higher measured and modelled $\beta$ resulting from lower $g_s$ as described in our
concurrent paper on modelling WTD effects on surface energy exchange over PDPSF (Mezbahuddin et al., 2014). Reductions in GPP resultant of plant water stress can also be corroborated by reductions in latent heat fluxes and consequent increases in $R_e$ during the late dry seasons as reported by Hirano et al. (2005) for our study site.

Deeper peat respiration during the deep WTD hydroperiod greatly increased due to abundant [O$_2$] in the deeper aerobic zone (Fig. 11) and resulting rapid microbial O$_2$ uptake (A17), C oxidation (A1, A2), growth (A25) and $R_h$ (A13, A20) as discussed in Sect. 2.1.2. This increase in deeper $R_h$ was greater than the reduction in surface and near-surface microbial respiration due to desiccation as discussed in Sect. 4.1.2, which led to a net increase in $R_e$ during this hydroperiod. This modelling hypothesis was further corroborated by larger soil CO$_2$ effluxes measured by Sundari et al. (2012) over our study site during the deep WTD hydroperiod than during the rest of the year. Cai et al. (2010) also measured a stimulation of respiration over a northern boreal peatland with deepening of WTD.

Deep WTD during the late dry season thus caused lower NEP (Figs. 2–6) by a combination of reduced GPP and increased $R_e$ (Figs. 7–8). This trend of decreased GPP due to plant water stress and increased $R_e$ due to enhanced deeper peat respiration during deep WTD hydroperiods was well corroborated by EC-gap filled CO$_2$ fluxes (Fig. 7) (Table 2) as well as by other field measurements over same site or similar ecosystems and hence validated our hypothesis 3.

4.2 Modelling WTD effects on annual tropical peatland C balance

4.2.1 Differences in annual EC-derived vs. modelled NEP, $R_e$ and GPP

Modelled annual NEP was considerably lower than the EC-gap filled annual NEP in 2003 and 2004 (Table 3). These lower NEP estimates were mainly attributed to larger modelled vs. gap-filled $R_e$ predominantly in the rainy seasons (Fig. 1) which was consistent throughout the study period and yielded larger slopes from modelled vs. gap-filled net CO$_2$ flux regressions (Table 2). During the rainy seasons throughout the study period, modelled water table never rose above the hollow surface leaving the top 0.05 m of hollow and the entire hummock unsaturated even when the WTD was the shallowest (Figs. 2–5). This trend was consistent with the observed daily WTDs (Figs. 4–5). Total porosity for the top 0.2 m of the modelled peat was 0.89 calculated from bulk density provided to the model (Mezbahuddin et al., 2014), consistent with field measurements in similar drained tropical peatlands (Couwenberg and Hooijer, 2013; Hooijer et al., 2012; Jauhiainen et al., 2012b; Takakai et al., 2006). Therefore, when $\theta$ in the top 0.2 m rose from $\sim 0.22$ m$^3$ m$^{-3}$ in the dry seasons to just above $0.30$ m$^3$ m$^{-3}$ in the rainy seasons, near surface peat in the model still had enough air filled porosity for soil respiration not to be suppressed (Fig. 1). Moreover, wet surface residue from frequent rainfall caused large flushes of modelled residue CO$_2$ effluxes during rainy seasons which were also apparent in soil respiration measurements by Sundari et al. (2012) at our study site. However, despite a similar rise in $\theta$ measured vs. modelled in the top 0.2 m, gap-filled CO$_2$ effluxes were much smaller than the modelled CO$_2$ effluxes during the rainy season (Fig. 1). Such smaller gap-filled vs. modelled nighttime NEP (Fig. 1) was consistent throughout the rainy seasons (November–April) of 2002–2005 as indicated by a slope of 1.24 from a regression of modelled on gap-filled net CO$_2$ fluxes ($n = 13218$). However, a slope of 0.98 from regression of modelled on EC-measured CO$_2$ fluxes ($n = 4464$) for the same hydroperiods indicated better model agreement with EC-measured fluxes. Since 75% of the total hourly CO$_2$ fluxes during rainy seasons of 2002–2005 were gap-filled, larger modelled vs. gap-filled CO$_2$ fluxes during these hydroperiods could largely contribute to larger modelled vs. EC-derived annual $R_e$ and hence lower modelled vs. EC-gap filled annual NEP estimates (Table 3).

Systematic uncertainties embedded in EC methodology were also thought to contribute to larger modelled vs. EC-derived monthly and annual $R_e$ estimates (Fig. 8) (Table 3). Nighttime EC NEP decreased with $u^*$ below 0.3 m s$^{-1}$ in our study site (Hirano et al., 2007) indicating the dependence of nighttime CO$_2$ flux measurements on above-canopy turbulent mixing (Miller et al., 2004). The measured CO$_2$ by plant and microbial respiration was independent of $u^*$ in the model. Possible underestimation of nighttime NEP resultant of uncertainty related to low $u^*$ threshold can be as large as $\sim 45\%$ of nighttime CO$_2$ fluxes estimated by Miller et al. (2004) for an Amazonian rainforest which would further contribute to larger modelled vs. EC-derived $R_e$ estimates.

Larger modelled vs. gap-filled $R_e$ contributed to larger modelled vs. gap-filled annual GPP (Fig. 8) (Table 3). In EC datasets, GPP was derived from extrapolated daytime $R_e$ (Sect. 2.2.2) and hence smaller gap-filled vs. modelled nighttime $R_e$ would cause smaller EC-derived GPP. A further cause of smaller EC-derived vs. modelled GPP could have been the incomplete ($\sim 80\%$) energy balance closure in EC measurements (Wilson et al., 2002) vs. complete energy balance closure in the model, which would reduce EC-derived ET and also possibly GPP (Table 3).

Apart from above mentioned systematic errors, random errors in EC measurements and gap filling could also contribute to divergence between modelled and EC-gap filled annual NEP estimates. Hirano et al. (2012) estimated an uncertainty of about $\pm 25$ g C m$^{-2}$ yr$^{-1}$ in NEP estimates resultant of random errors in EC measurements and gap filling during 2002–2008 over our study site. Continuous data gaps due to instrumental failure can induce additional uncertainty in EC-gap filled annual NEP estimates. Richardson and Hollinger (2007) estimated an additional uncertainty of
±30 g C m$^{-2}$ yr$^{-1}$ in NEP estimates for a week long gap in flux data over a range of forested FLUXNET sites. Additional uncertainties in EC-gap filled NEP estimates due to errors in filling continuous data gaps, however, were not quantified for the EC-gap filled NEP datasets used in our study. Such uncertainties could be substantial in EC-gap filled NEP estimates used in our study due to the presence of very long continuous data gaps (e.g. 11–22 November 2002, 3–31 January 2003, 29 March–22 May 2003, 24 June–26 July 2003 and 14 September to 9 October 2004) resultant of instrumental failures. Positive uncertainties due to above mentioned random errors could further contribute to larger EC-gap filled vs. modelled annual NEP, and hence smaller EC-derived vs. modelled annual $R_e$ and GPP estimates.

All of these above mentioned sources of larger modelled vs. EC-derived $R_e$ and GPP estimates were related to EC methodology and gap filling. These discrepancies between modelled and EC-derived $R_e$ and GPP aggregates, however, could not be resolved in our modelling since, unlike EC datasets, every single mole of CO$_2$ that was modelled from fundamental ecosystem processes was counted in the modelled C budget.

4.2.2 Differences in annual WTD effects on EC-derived vs. modelled GPP and $R_e$

At an annual time scale, reductions in both GPP and $R_e$ with a gradually shallower WTD from 2002 to 2005 in ecosystems were not corroborated by changes in GPP and $R_e$ partitioned from EC-gap filled net CO$_2$ fluxes by Hirano et al. (2007) (Sect. 2.2.2) (Table 3) although monthly partitioned GPP and $R_e$ showed the similar seasonal trends as those modelled (Fig. 8). Accumulation of above-mentioned discrepancies between modelled and EC-gap filled fluxes (Sect. 4.2.1) over a larger time scale (e.g. from monthly to annual) obscured the agreement between the modelled and EC-gap filled trends in WTD effects on annual GPP and $R_e$. Furthermore, as opposed to EC-derived $R_e$ that was used to calculate EC-derived GPP, modelled $R_e$ ($R_a + R_h$) was driven by modelled GPP thereby contributing to deviation between WTD effects on modelled vs. EC-derived annual GPP and $R_e$. Modelled $R_a$ was directly dependent on fixed C products during photosynthesis. Modelled $R_h$ was also dependent on fixed C products in a diurnal time scale through root exudates as well as in a seasonal time scale through above and below ground litter fall.

Difference in flux partitioning methods may also produce variable agreements between modelled and EC-derived annual GPP and $R_e$ aggregates. Desai et al. (2008) found a variability of ∼10 % in annual GPP and $R_e$ estimates for 23 different flux partitioning methods over six European temperate forest. This difference was comparable to typical interannual variability in GPP and $R_e$ over those sites. They also reported higher variability among those partitioning methods in estimating nighttime $R_e$ that arose from frequent data gaps during calm night hours. Annual GPP and $R_e$ aggregates for different partitioning methods in their study also varied abruptly for a Mediterranean site indicating spatial variability of partitioning method induced variation in annual GPP and $R_e$ estimates. Moreover, artificially created data gaps of 10 % of total half-hourly net CO$_2$ fluxes in a year imparted additional 6–7 % variability in annual GPP and $R_e$ aggregates in their study. These artificially created data gaps also contributed to a variability of ±25 g C m$^{-2}$ yr$^{-1}$ in annual NEP estimated for 15 different gap filling techniques (Moffat et al., 2007). Above mentioned partitioning method induced variations in annual GPP and $R_e$ aggregates as estimated by Desai et al. (2008) could be more substantial over the drainage affected tropical peat swamp forest under this study due to longer growing season (i.e. tropical evergreen vs. temperate winter deciduous phenology) and larger soil CO$_2$ effluxes (i.e. tropical drained peat soils vs. temperate mineral soils).

We did not, however, examine whether a partitioning method other than the one used in this study (Sect. 2.2.2) could produce better agreement between modelled and EC-derived annual GPP and $R_e$ aggregates. Inclusion of such a substantial study could obscure our main objective of portraying mechanistic relationships between seasonal variation in WTD and that in NEP which has been adequately corroborated by the regression analyses between hourly modelled vs. EC measured net ecosystem CO$_2$ fluxes (Table 2). However, comparing different flux partitioning methods along with sophisticated ecosystem modelling is likely to direct a separate future research that would offer increased confidence in annual GPP and $R_e$ estimates over tropical peatland ecosystems.

Despite the above mentioned divergence between modelled vs. EC-derived $R_e$ and GPP estimates, components of modelled annual C balance were comparable with biometric measurements and estimations from other studies on similar ecosystems. Modelled net primary productivity (NPP) was comparable with values estimated for Amazonian rainforests and oil palm plantations in tropical peatlands (Table 3). Modelled $R_h$ was comparable with $R_h$ measured for a mature Acacia plantation on drained Indonesian peatlands (Table 3).

4.3 Effects of WTD on modelled annual C balance in drained vs. undrained simulation

Reduced WTD by an average of ∼0.5 m resulted in an increased mean NEP by 270 g C m$^{-2}$ yr$^{-1}$ in undrained simulation compared to that in drained simulation over 2002–2005 (Table 4). This modelled trend was corroborated by measurements of Hirano et al. (2012) who found a larger mean NEP by ∼153 g C m$^{-2}$ yr$^{-1}$ in a nearby undrained peat swamp forest than that in our drainage affected peat swamp forest site over 2004–2008. This increase in NEP in undrained simulation was modelled through greater suppressions of $R_a$ and $R_h$ than of GPP with reduced WTD (Table 4).
Apart from net vertical CO₂ exchange, drainage of tropical peatlands can also trigger substantial C losses through lateral transport of dissolved organic C (DOC). Moore et al. (2013) measured a drainage-induced additional C loss of 20 g C m⁻² yr⁻¹ through the export of DOC in our study area. We, however, simulated a negligible additional increase in C losses through export of DOC (1–2 g C m⁻² yr⁻¹) due to drainage in this modelling study. Since transport of DOC heavily depends upon total amount of catchment discharge (Moore et al., 2013) this discrepancy between our point scale study and their watershed scale measurements on DOC transport is reasonable. However, up scaling our modelling to watershed scale might be a potential opportunity to examine the effects of drainage on C losses through export of DOC in drained tropical peatlands.

Even in the undrained simulation, NEP for 2002–2004 indicated substantial C losses from PDPSF (Table 4). Similarly large C losses were estimated from EC-gap filled NEP during 2004–2007 by Hirano et al. (2012) over a nearby similar undrained peat swamp forest. They also predicted from a simple linear regression analysis of NEP on WTD that maintaining a mean annual WTD within 0.03 m below the hollow surface could bring the undrained peatland ecosystem to C neutrality. In line with their simple prediction, a much more sophisticated process-based undrained simulation by ecosys in our study resulted in a near C neutral NEP during the wettest year 2005 with a mean annual WTD of 0.10 m above the hollow surface (Table 4).

C losses modelled in the undrained simulation and observed in the undrained peat swamp forest, however, may be a recent phenomenon since a long term apparent C accumulation study showed that Central Kalimantan peatlands have been accumulating C at a rate of 31 g C m⁻² yr⁻¹ for last ∼12 000 yr (Dommain et al., 2011) and that the peatlands around our study site have been accumulating C at a rate of 56 g C m⁻² yr⁻¹ for last ∼20 000 yr (Page et al., 2004). One of the main reasons behind the modelled negative NEP even in the undrained simulation may be that the precipitation within our study period (2002–2005) was less than the long-term average, which led to deeper WTD than the long-term mean. The wettest year in our study period (2005) experienced a total annual precipitation which was considered as “normal” annual precipitation (2570 mm yr⁻¹) measured over Indonesian Borneo during 1994–2004 (Takahashi et al., 2004). The year 1999 was the wettest within their measurement period with an annual precipitation of 3788 mm that caused the water table to remain above the ground throughout the year (Wösten et al., 2008). This speculation of large C losses due to reduced precipitation and consequent deeper WTD in recent years can be further corroborated by the cessation of Central Kalimantan peat growth during last ∼5000 yr as reported by Dommain et al. (2011) due to WTD drawdowns caused by increased El Niño frequency and intensity.

5 Conclusions

Ecosys successfully simulated the reduction of tropical peatland NEP during shallow and deep WTD hydroperiods with respect to those in intermediate WTD hydroperiods for four years, i.e., 2002–2005 over PDPSF (Figs. 2–5) (Table 2). Reduction of NEP during shallow WTD was modelled mainly through reduced GPP and that during deep WTD hydroperiods was modelled by a combination of reduced GPP and increased Rₑ (Figs. 6–8). Seasonal variation in NEP that was apparent in the measurements was thought to be caused by the following key responses that were modelled using following algorithms of WTD effects on GPP and Rₑ from basic independent research fed by site specific inputs (Sect. 2.2.3):

1. Shallow WTD during rainy seasons reduced modelled NEP by explicitly simulating slower convective-dispersive O₂ transport through soils and roots and hence slower root O₂ uptake, slower soil nutrient (phosphorus) transformations (A26), slower root nutrient (phosphorus) uptake and growth (C23), and consequently lower leaf nutrient (phosphorus) status (C11) and slower CO₂ fixation (C6) (Sect. 4.1.1).

2. WTD drawdown during early and late dry seasons in the model increased deeper peat respiration due to better aeration by explicitly simulating higher [O₂] (A17) from more rapid O₂ transport through soils, and hence more rapid microbial and root oxidation-reduction reactions (A3, A5), greater microbial O₂ uptake and energy yields (A20) driving more rapid microbial growth (A25) and respiration (A13, A14, A20) (Sects. 4.1.2 and 4.1.3).

3. Deeper WTD during late dry seasons in the model reduced NEP through plant water stress by explicitly simulating declines in gₛ and their effects on CO₂ fixation from hydraulically driven water transport along soil-plant-atmosphere water potential gradients (B1–B14) (Sect. 4.1.3).

Though our modelling effort reasonably well simulated seasonal WTD effects on NEP over PDPSF, the modelled GPP and Rₑ aggregates were systematically larger than the EC-derived estimates (Fig. 8) (Table 3). The possible reasons for these discrepancies are discussed in Sect. 4.2. Despite the modelled vs. EC-derived divergence in monthly and annual Rₑ and GPP aggregates, this study showed for the first time the application of a detailed process based model in gaining insights on nonlinear relationships between WTD and net ecosystem CO₂ exchange of tropical peatlands, as recommended by Murdiyarso et al. (2010). The findings of this study showed that the duration and intensity of the dry season with deeper WTD had profound effects on tropical peatland CO₂ emissions irrespective of disturbance (e.g. drainage). This has an important implication in terms of the fate of
tropical peatland C storage under future climate change scenarios since Li et al. (2007) predicted a general drying trend and consequent WTD drawdown over Southeast Asian peatlands during this century using 11 land surface models. Moreover, the response of tropical peatland CO$_2$ exchange to disturbance (e.g. drainage) was also investigated by our model sensitivity test for drained vs. undrained condition. Insights gained from our modelling effort thus can improve our predictive capacity for the effects of WTD fluctuations arising from interactions between seasonality in precipitation and artificial drainage on tropical peatland C balance.

Our point scale modelling reasonably delineated WTD effects on NEP over a homogeneous patch of drainage affected tropical peat swamp forest in terms of plant functional type (PFT) (i.e. tropical evergreen vascular vegetation) and land use (i.e. drained forest). However, this modelling can be up scaled to an ecosystem level by model inputs of weather data (Sect. 2.2.3); soil physical, hydrological, chemical and biological properties (Sect. 2.2.3) (Table 1) (Mezbahuddin et al., 2014); PFT (e.g. moss vs. vascular plants, evergreen vs. deciduous); and disturbance (e.g. drainage as discussed in Sect. 2.2.6, fire, logging etc.) representing a particular peatland ecosystem. Such up scaling studies over tropical peatlands could provide us with improved predictive capacity on management opportunities (e.g. undrained vs. drained, reforestation/afforestation vs. deforestation, unburnt vs. burnt) for reducing C emissions. This capacity might be very important for planning long-term tropical peatland rehabilitation projects and mapping peat C sequestration for current REDD+ (Reducing Emissions from Deforestation and Forest Degradation) scheme.

Supplementary material related to this article is available online at http://www.biogeosciences.net/11/577/2014/bg-11-577-2014-supplement.pdf.

Acknowledgements. Computing facilities for ecosys were provided by University of Alberta and Compute Canada. Modelling was funded by Natural Sciences and Engineering Research Council of Canada, and Faculty of Graduate Studies and Research, University of Alberta. Field data collection was partly supported by JST/JICA SATREPS Project. We also thank AsiaFlux for providing a platform for sharing the field data used in this study.

Edited by: V. Brovkin

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