

*Vulnerability to drought and
soil carbon exchange of valley
forest in Central Amazonia
(Brazil)*

Fabrício Berton Zanchi

Cover page: The hostile ZF2 forest

Back cover: Images aligned horizontally, left-to-right side of the page: Beginning of my PhD. life, very first days in Amsterdam; after 11 months, I was back in Manaus - Brazil to do the fieldwork in the outstanding Cuieiras Reserve (tower landscape view); sap flow system installation; ZF2 tower; soil respiration system; Pasture; Drainage experiment (tube installation); Campina forest; Campina soil profile; Manuella's fieldwork dog; fieldwork lodge and then, the last days of the Thesis editing (hairless).

Key words: Amazon forest / heath forest / soil respiration / analytical model / biogeochemistry / hydrological cycle / climate change / carbon cycle / pasture biomass / LAI

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Printed by: Centrale Huisdrukkerij van de Vrije Universiteit, Amsterdam.

VRIJE UNIVERSITEIT

Vulnerability to drought and soil carbon exchange of valley forest in Central Amazonia (Brazil)

ACADEMISCH PROEFSCHRIFT

ter verkrijging van de graad Doctor aan
de Vrije Universiteit Amsterdam,
op gezag van de rector magnificus
prof.dr. F.A. van der Duyn Schouten,
in het openbaar te verdedigen
ten overstaan van de promotiecommissie
van de Faculteit der Aard- en Levenswetenschappen
op woensdag 18 december 2013 om 9.45 uur
in de aula van de universiteit,
De Boelelaan 1105

door

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geboren te Umuarama, Brazilië

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Chapter 1

Introduction

1.1 The Amazon region and development

The Amazon region contains more than half of Earth's remaining tropical forest and is the largest contiguous tropical forest ecosystem in the world (6.6 million km²). 80% of the region is still in a pristine state (INPE, 2008), 5.4 million km² is located in Brazil with the remaining areas in Bolivia, Colombia, Ecuador, Guyana, Peru, Surinam and Venezuela (Houghton et al., 1995; IBGE, 1997; Waterloo et al., 2006).

The region is a mosaic of landscapes, formed in the Tertiary and Quaternary geological period and comprises at present 13% cattle pasture and agricultural lands, concentrated around the Brazilian states of Acre, Rondônia, Mato Grosso, Tocantins and Pará (IBGE, 1997; Soares-Filho et al., 2006). 6% consist of savannah-like vegetation (*Cerrado*, *Campina* and *Campinarana*) and is located in areas around Roraima and the North and South of Amazonas State (McClain et al., 1997; IBGE, 1997; Luizão et al., 2007), 70% is evergreen forest (*terra-firme* forest) and can be found throughout the whole Amazon region including areas of swamp, *Igapó* and *Várzea* forests located near along the large rivers (Prance and Schubart, 1978; Klinge and Medina, 1979; Prance, 1979; Anderson, 1981; Luizão et al., 2007).

Before 1960 the population of the Amazon region settled primarily along the main large rivers due to restricted transportation access. At the time of the Brazilian military government, the Amazon region was opened to exploration by the construction of railways and roads that increased access in the rain forest, such as Belém–Brasília (BR–153), Cuiabá–Porto Velho (BR–364), Transamazonica road (North–Northwest) and Cuiabá–Santarém (BR–163). Today these areas represent the arch of the deforestation (arco do desmatamento), coinciding with the Amazonian agricultural frontier (Malhi et al., 2008; Carneiro-Filho and Souza, 2009). The Amazon region is usually considered to have a low capacity to recover from externally imposed perturbations

(resilience) such as these road constructions, government colonization programs, and large-scale cattle ranching (Carvalho et al., 2001; Nepstad et al., 2002a; Soares-Filho et al., 2004; Nepstad et al., 2008). Beyond the pressures of these forms of deforestation there are also others including mining, hydroelectric power constructions, timber exploration, indigenous land protection and petroleum exploration (Carneiro-Filho and Souza, 2009) (Figure 1.1). All of these pressures combined represent 75% of Brazilian CO₂ emissions (Houghton et al., 2000).

Deforestation in the Amazon region occurred because the newly paved roads ensured not only increasing regional deforestation rates, but also initiated new frontiers for large-scale occupation, (Nepstad et al., 2008) as the roads became economic focal points that further stimulated deforestation (Soares-Filho et al., 2004). While deforestation generally begins with timber exploitation, and selective logging does not cause an immediate land use change *per se*, it does lead to large scale deforestation through development of new unofficial roads and the arrival of colonist farmers and loggers that subsequently occupy and clear the forest (Davidson et al., 2012). The first colonists work on the land for about 10 years and then abandon the area due to declining soil fertility (Fearnside, 1993) and high expenses associated with fertilizing and recovering the soil after applying the poor clearing-fire methods that are unfortunately still common practice in the Amazon region. It is difficult to control and reduce the impacts on remaining forest, because the international and national demands for cattle and livestock feed maintain the pressure for further deforestation in the Amazon rain forest (Nepstad et al., 2008; Davidson et al., 2012). While cattle pasture remains the dominant land use type for cleared land, the growing importance of larger and faster conversion to cropland – mostly for soybean export – has caused a trend of forest loss in Amazonia since the early 2000s (Davidson et al., 2012). The total deforestation over the past decades has been about 750,000 km² of the Brazilian Amazon only (INPE, 2008; Nobre and Borma, 2009). The Brazilian government has chosen for a strategy of expansion of protected areas, while imposing physical barriers and a fiscalisation policy. This strategy is now paying off, reducing deforestation (Carneiro-Filho and Souza, 2009; INPE, 2011; Davidson et al., 2012). The associated measures help to maintain larger areas of forest with restricted accessibility in the edge of the arch of deforestation and have curbed deforestation through a steep decline in forest clearing in the Brazilian Amazon, from nearly 28,000 km² in 2004, 11.968 km² between 2007–2008 and to less than 7,000 km² in 2011 (INPE, 2008, 2011).

To understand the possible future land use change and the interaction with climate change in the Amazon region, Soares-Filho et al. (2004) and Nepstad et al. (2008) used two future scenarios, which influence the carbon and hydrological cycles and the economic regulation of the Amazon region. The first is a *business-as-usual policy* scenario, which covers the response of deforestation and land abandonment to road construction through estimates based of historical rates of Amazon regions. The second scenario is the *governance scenario*

where several plausible improvements in the enforcement of environmental regulations, support for sustainable land-use systems, and local institutional capacity are invoked to modify the historical rates. These scenarios prescribe two possible types of occupation in the Amazon region. The expected trend of deforestation within the business-as-usual scenario (Fig. 1.1) would lead, in the middle of this century, to a reduction of about 40% in the current 5.4 million km² of forests in the Amazon basin. With the governance scenario, deforestation can decrease by 55% compared to the business-as-usual (Soares-Filho et al., 2005, 2006; Nepstad et al., 2008).

The business-as-usual scenario as shown in Fig. 1.1 will encourage further expansion of agriculture and timber exploitation which may lead to a massive conversion of forests into pastures and agricultural areas, and consequently, profound loss of still unknown genetic heritage of various ecosystems of the Amazon. We now turn our attention to the possible impacts of these pressures on the water and carbon cycle of Amazonian forests.

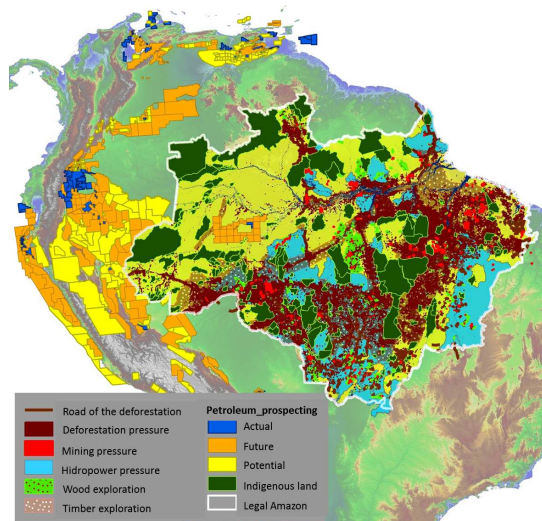


Figure 1.1: Set of pressures and threats on lands in the Brazilian Amazon region. Adapted from (Carneiro-Filho and Souza, 2009)

1.1.1 Amazonian forests and the water and carbon cycle

The net carbon exchange of a forest is the balance of the carbon uptake by the forest through photosynthesis and subsequent release by plant and soil respiration and decomposition of organic matter, root exudation and leaching in runoff (Malhi et al., 2009a). Most global circulation models (GCMs) pre-

dict that greenhouse gas accumulation and associated increases in the radiative forcing of the atmosphere will result in a decline of rainfall in eastern Amazonia by about 20% towards the end of the century (IPCC, 2007). The net carbon stored within the Amazon ecosystem is to first order the difference between carbon uptake during photosynthesis and carbon loss during respiration (Fig. 1.2a)). Thus net ecosystem production (NEP) is calculated as:

$$\text{NEP} = R_{Het} - (\text{GPP} - R_{Aut}) \quad (1.1)$$

where, a negative flux indicates a net carbon uptake. Gross primary production (GPP) is the rate at which an ecosystem captures and stores a given amount of chemical energy as biomass over a given length of time – in other words, the carbon uptake during photosynthesis. R_{Aut} , the autotrophic respiration, is the carbon loss from growth and maintenance respiration by plants and R_{Het} , the heterotrophic respiration, is the loss of carbon by soil micro-organisms as they decompose organic matter. The excess of carbon uptake minus autotrophic respiration by plants can be used to grow new biomass and this difference ($\text{GPP} - R_{Aut}$) is called the net primary production (NPP) (Eq. 1.2):

$$\text{NPP} = (\text{GPP} - R_{Aut}) \quad (1.2)$$

Saatchi et al. (2011) and Baccini et al. (2012) estimated the total Amazon Basin carbon (C), based on a large number of biomass plot data from many different forest types distributed over the Amazon Basin using remote sensing to scale this up to larger areas. The total Amazon Basin above-ground forest biomass was estimated at 108.3 Gt C (Baccini et al., 2012). Including live below-ground biomass this increased to 121 Gt C (Saatchi et al., 2011). Dixon et al. (1994), Bonan (2008) and Saatchi et al. (2011) estimated that the Amazon region has a high rate of primary production, amounting to about 30-50% of the global average.

The estimated total stock of carbon in the old growth *terra-firme* forest is still uncertain, some authors found values which vary from 81.9 to 91.2 Gt C (Saatchi et al., 2011; Baccini et al., 2012). Floodplain inundated forests store 44 Mg C ha⁻¹, woodland savanna 31 Mg C ha⁻¹ and open shrubs 23 Mg C ha⁻¹ (Baccini et al., 2012). The net effect of Amazonian deforestation and reforestation results in an annual net C source of 0.15-0.35 Pg C, and the total forest fire C emission is about 0.23–0.54 Pg C yr⁻¹ (van der Werf, 2006; Davidson et al., 2012). The combined carbon emissions from fire and from logging would annually release 0.2-0.8 Pg C. The Amazonian rivers are supersaturated with dissolved CO₂, which may eventually release to the atmosphere an estimated 0.5 Pg C yr⁻¹ (Richey et al., 2002). There exists considerable uncertainty about the estimates of the carbon balance across the Amazonian ecosystem (NEP). According to (Phillips et al., 2009), mature forests are accumulating on average about 0.4 Pg C yr⁻¹. The uncertainties of the total carbon stock estimates will eventually be reduced when field inventories and

the geographical distance in the Amazon region match and provide more reliable data sets to improve the remote sensing techniques for determining local carbon stock.

The hydrological cycle is also important to the whole Amazonian ecosystem, in the first instance because of drought effects, while the spatial variation in rainfall may also partly control and limit forest productivity across the Amazon Basin (Cox et al., 2004; Phillips et al., 2009; Lewis et al., 2011; Hirota et al., 2011). The change in the water balance Wb over time t is formulated as:

$$\frac{dWb}{dt} = P - ET - R + I - D \quad (1.3)$$

where, P is the precipitation, ET is evapotranspiration, R is runoff, I is infiltration into the soil and D drainage to groundwater, all expressed in mm. Many studies have been conducted to quantify the water balance in the Amazon region (*i.e.*, Herrera (1979), Lloyd et al. (1988), Ubarana (1996), Marin et al. (2000), Cuartas et al. (2007), Tomasella et al. (2007), Costa et al. (2010) and Germer et al. (2010)). The water exchange also makes the region globally relevant because the Amazon River system produces about 20% of the World's freshwater discharge (Salati and Vose, 1984). This amount corresponds to approximately eight trillion tons of water evaporating from the Amazon forests each year. This volume of water has important impacts on the global atmospheric circulation, of which 3.8 trillion of $\text{m}^3 \text{yr}^{-1}$ is transported from the Amazon region to the South American region (Fearnside, 2004). This in turn represents 52% of the total discharge of the Amazon River (Fig. 1.2b)).

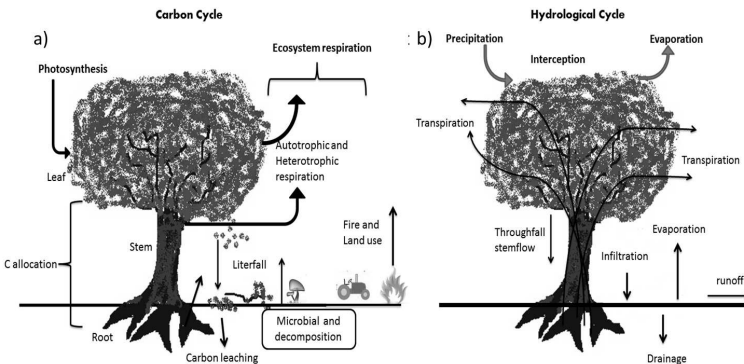


Figure 1.2: a) Amazonian forest carbon cycle and b) hydrological cycle flows. Redrawn from Bonan (2008).

Other Brazilian regions and South America as a whole depend on the water vapour coming from Amazonia for their agriculture, hydropower dams and water supply Fearnside (2004), Soares and Marengo (2009) and Marengo et al.

(2012). Part of Brazil suffered from energy and water supply rationing in 2001, all because of low water vapour feedback from the low level jets from the Amazon region to those regions. Some climatologists argue that these anomalies will become more frequent (Marengo, 2004) as greenhouse gases accumulate further in the atmosphere (Oyama and Nobre, 2003; Hansen et al., 2006). This is critically important for ecosystem health as modelling studies indicate that the ecosystem fluxes depend strongly on the availability of water for plants (Nobre and Borma, 2009; Hirota et al., 2011).

1.1.2 Amazonian forest – environment interactions

Apart from the immediate threats of deforestation, there is increasing evidence that global climate change, through reduced rainfall and increased temperature, can have direct and long-term effects on the functioning and sustainability of Amazonian ecosystems (Oyama and Nobre, 2003; Cox et al., 2004; Phillips et al., 2009; Lewis et al., 2011). Those predicted changes in climate may be further amplified regionally by progressing deforestation in the Amazon basin because recycling of water in the Amazon Basin itself generates a significant fraction of its rainfall (Aragão, 2012; Spracklen et al., 2012). The resulting change in the water and carbon cycles of the Amazon may have potentially dangerous positive feedback effects through decreasing biomass production and reduced evaporation and increased soil CO₂ emissions, thereby exacerbating changes in global climate (Cox et al., 2000; Huntingford et al., 2013). The probability that this may happen in the Amazon is real, as the region has already been shown to be very sensitive to changes in rainfall (Nepstad et al. (2008) and Costa et al. (2010)). However and importantly, these drought-biomass interactions are not well understood. Recent modelling results indicate that the full die-back scenarios are less likely, but that uncertainties in the effects of CO₂ fertilization play a major role (Cox et al., 2013; Huntingford et al., 2013). Following the severe droughts in 2005 and 2010, forced by variability in the Atlantic Multi-decadal Oscillation (Marengo, 2004; Marengo et al., 2012), new studies have provided important information on the effects of water stress on Amazon forests (*e.g.* Phillips et al. (2010)). Such droughts appear to become more frequent in the Amazon region (Marengo, 2004) and may increase Amazonian forest fire vulnerability (Aragão et al., 2008).

Amazonian ecosystems can allocate carbon at high rates to the foliage ensuring more light and CO₂ absorption (Dixon et al., 1994; Bonan, 2008). On the other hand, inefficient CO₂ foliage allocations exist if there are not sufficient nutrients and water available (Norby et al., 2010) to support gross primary production. The combination of access to deep soil water and less cloudiness in the dry season permits continued plant photosynthesis throughout most of the dry season (Saleska et al., 2003; Rocha et al., 2009; Davidson et al., 2012). At a general level, these studies thus suggest that Amazonian forest is resilient to seasonal and moderate drought, and also resilient to initial disturbances (Saleska et al., 2003; Huete et al., 2006; Lewis et al., 2011). However,

repeated or prolonged disturbances may change the forest structure (land-use) and nutrient dynamics and lead to long-term changes in vegetation composition and carbon loss. Specific drought manipulation experiments have been conducted in the Amazon region, near Santarém, Brazil (SECAFLORESTA) (Nepstad et al., 2002a) and West of Belém in Caxiuanã (ESECAFLOR, (Costa et al., 2010)) with 8 and over 10 years of results available, respectively. These show strong ecosystem water dependency and high carbon loss, especially after prolonged drought. The evidence resulting from natural droughts and the drought manipulation experiments indicate a similar trend of mortality in response to the intensity of the dry season (Phillips et al., 2010). As mentioned by Lewis et al. (2011), two major droughts in recent decades showed that these losses may largely offset the net gains of 0.4 Pg C yr^{-1} in intact Amazon forest above-ground biomass in non-drought years.

Dry-season water stress is likely to become more important in East Amazonia over the 21st century (Malhi et al., 2009b) as the region tends towards a climate that is more appropriate to savannah than to a seasonal forest. Because of the many pressures from land-use change (Nepstad et al., 2006) these seasonal forests may still be resilient to seasonal drought but are likely to face intensified water stress caused by the higher temperatures. They may also be more vulnerable to fires, especially those that are of natural origin, which are currently still rare in much of Amazonia, while those of human origin more frequent.

Although diverse events can trigger shifts in vegetation, recent studies show that a loss of resilience can also pave the way to an abrupt change into a different, alternative state of the vegetation. For example, Nobre and Borma (2009) and Hirota et al. (2011) show that Amazon forests can be vulnerable to such regime shifts. These shifts are caused by the ecosystem response to gradually changing conditions of annual precipitation, climate change, nutrient loading and habitat fragmentation. This suggests that strategies for sustainable management of such ecosystems should focus on maintaining resilience. Otherwise, repeated land-use pressure and drought may have important decadal-scale impacts on the Amazonian and global carbon cycle (Fig. 1.1) and turn the Amazon from a likely sink into a source of carbon to the atmosphere.

Most experimental and model studies assume that the rain forest of the dominant, well-drained upland *terra-firme* forest growing on clayey to loamy soils (Oxisols, Ultisols), has a homogeneous cover (Cox et al., 2000; Oyama and Nobre, 2003; Cox et al., 2004). It is however also important to study the other components of the system such as the pastures of south-western Amazonia and the central Amazonian rain forest landscapes (Fig. 1.3). In these a different mosaic of vegetation types is found, consisting of tall *terra-firme* evergreen forest, and seasonally inundated *Várzea* or *Igapó* valley forests along the major rivers. Campinarana ecotone areas and riparian valley forests occur along the smaller rivers. These valley forest areas occupy about 35% percent of central Amazonian landscape (Nobre et al., 2011). More than 30% of the total groundwater recharge comes from the slope and plateau forests of these

areas (Waterloo et al., 2006). Low-statured Campina forests *sensu* (Anderson et al., 1975) reflect topography-induced variations in strongly leached sandy soils with poor nutrient and drainage conditions covering about 6% of the Amazon region (Prance and Schubart, 1978). These Campina forests, whether of natural or human-induced origin, as well as pastures in the south-west with their particular characteristics, may reduce the overall resilience of the system.

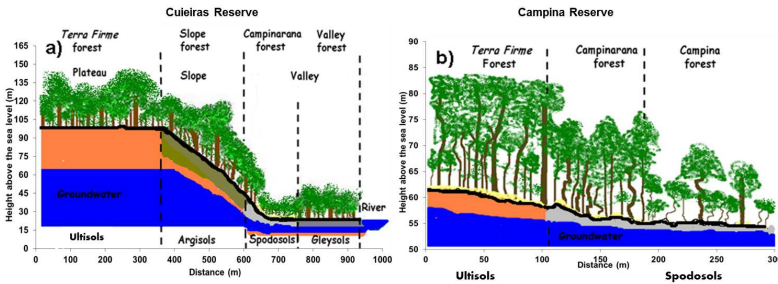


Figure 1.3: Vegetation and soil type differences of toposequences of a) the Cuieiras Reserve and b) the Campina Reserve. After Chauvel et al. (1987).

Thus, valley forest may suffer from such a *Campinarana* type reduction (deeper water table) if the prediction of global circulation models (GCMs) of 20% less rainfall would become true in the Amazon Basin (Cox et al., 2000; Oyama and Nobre, 2003; Cox et al., 2004). As groundwater recharge on the slope and plateau – which maintains a steady flow of groundwater towards the valley – would decrease under such drying conditions, this could trigger the valley areas into a much dryer state than they currently occupy in the vegetation – climate system. This prospect increases the vulnerability of the valley forest considerably.

1.2 Objectives and thesis outline

In central Amazonia the climate does not show much seasonality. Permanently wet soils in the valley area are maintained by a continuous supply of groundwater from plateau and slope areas. Whilst Oxisols and Ultisols on the slopes and plateau have a significant clay fraction, valley soils generally consist of strongly leached quartz sand (Spodosols) with little capacity to retain water or nutrients (Chauvel et al., 1987). Nutrients may be retained in the valley soil by the soil carbon fraction (humus) and the high groundwater table, which may also buffer the pH to some extent, allowing tall forests to grow on these white sands (Guillaumet, 1987; Luizão, 1996). Under the present climate conditions (Marengo, 2004; Cox et al., 2004; Nobre and Borma, 2009) the vegetation

in poorly-drained valleys is not likely to experience significant drought and may even benefit from a slightly dryer climate as root water uptake, photosynthesis and decomposition may be reduced in saturated soils. However, if reduced recharge on the plateaus and slopes would result in a significant lowering of the valley groundwater table over extended periods, soil carbon could decompose and oxidise, leaving an extremely poor and potentially toxic soil. This might trigger a change from an existing tall valley forest to a heath forest with more sclerophyllic leaves. This forest is locally known as *Campinarana*, *Campina* or *Bana* forest (Anderson, 1981; Sobrado and Medina, 1980; Reich et al., 1994). The presence of *Campinarana* forest between the tall forest in the valley bottom and that on the slope (Waterloo et al., 2006) is an indication that a relatively small change in the valley groundwater level may be sufficient to cause an extension of the *Campinarana* zone towards the river. Although these *Campina* forests are quite common in the tropics, little is known about their hydrologic functioning (e.g. evapotranspiration rate) and their carbon balance (Becker, 1996b; Tyree et al., 1998).

Investigating this complexity can shed light on the question whether the valley vegetation may increase its production and evapotranspiration, or collapse and change to a *Campina* type of forest with presumably much reduced production and evaporation rates. The second possibility would mitigate the effects of the dying off of the forests on the plateau on the regional climate, whereas the first and third option represents a positive feedback where rainfall may even become further reduced.

To investigate these issues we conducted an experimental study in the Brazilian Amazon region. The research was performed across seven different ecosystems (cattle pasture, stunted heath forest (SHF), tall heath forest (THF) and a topo-sequence of riverine valley, *Campinarana*, Slope and Plateau forests) with all forests types having their own particular micro-environmental adaptation. The perspective of the study is to determine primarily the variability in water, carbon and nutrient exchange of the valley and of the *Campina* forest. Soil respiration, dissolved inorganic and organic carbon from the water and soil, water balance, Leaf Area Index (LAI) and the decomposition rates that occur along the topographic transect (Plateau-valley/bottom areas) were also determined. To provide a comparison and suggest plausible scenarios for the sensitivity of forests to drought we performed an artificial drainage experiment in the valley forest.

This study is part of the WOTRO study Vulnerability and resilience of the Brazilian Amazon forests and human environment to changes in land-use and climate (W.01. 65.217.00). This integrated project investigated several aspects of resilience and feed-back processes at a range of scales. The project comprised two other PhD studies and two post-doc positions. While this thesis studied small-scale aspects in the wet valley bottoms by comparing these to other forest patterns in the central Amazon, the second PhD. study investigated the fragmentation and hydrology at the scale of these fragments. In the third PhD study deforestation drivers were studied at the scale of municipi-

palties (a few square kilometers) to that of the state (the state of Rondônia). Resilience at basin-wide and international scale were investigated in the two post-doc projects. The overall study aimed to integrate the different scales through conceptual modelling (building system diagrams). It was hoped that in this way, the project would also be able to link to the feed-back of stakeholders, such as land owners.

The main objectives of the present study are thus to:

- i Improve the methodology for determining soil respiration in the central Amazonia to determine the relation between respiration and soil temperature;
- ii Determine the soil carbon and nutrient change from a saturated artificially drained valley soil compared to the heath forest from *Campina* reserve;
- iii Determine the hydrological and biogeochemical behavior of a *Campina* forest catchment and compare it with undisturbed *terra-firma* forest;
- iv Determine the seasonal and inter-seasonal variation of LAI from a cattle pasture in Rondônia state, Brazil and compare it with other ecosystems.

This thesis is organized in seven chapters. This first chapter serves as the introduction. The second, third, fourth, fifth and sixth chapters correspond to five articles published/submitted to peer-reviewed journals and to the objectives listed above. The seventh chapter is the synthesis and provides conclusions and recommendations. The second chapter described the development of the new methodology for determining soil respiration in central Amazonia. The third chapter covers the analysis of the soil respiration for all seven different rain forests. The new approach developed uses a new sinusoidal model to minimize the hysteresis between soil respiration and soil temperature. The fourth chapter presents the study of the meteorology, hydrology and biogeochemistry of a *Campina* reserve near Manaus, Brazil. The fifth chapter pertains to the groundwater manipulation experiment, and determines decomposition rates, biogeochemistry and soil respiration data. The sixth chapter presents six years of Leaf Area Index measurements from a cattle pasture in Rondônia state, Brazil. The behaviour of the cattle pasture is presented through the seasonal variation of the vegetation cover, groundwater table and micrometeorological data from the whole period of measurement. These results provide a long-term data set of cattle pasture vegetation dynamics, including the cattle presence for the six seasonal periods. Finally, the seventh chapter synthesizes the findings of the previous chapters and indicates future directions to estimate the biogeochemistry in relation to the water balance of the central Amazon with a view to arrive at better estimates of the resilience of the Amazon forests.

Chapter 2

Soil CO₂ efflux in Central Amazonia: Environmental and methodological effects¹

2.1 Introduction

Changes in terrestrial ecosystems have contributed to atmospheric CO₂ enrichment in both pre-industrial and industrial times. DeFries et al. (1999) reported that approximately 40% of the increase in CO₂ since 1800 can be attributed to land-use changes. Increases in atmospheric greenhouse gas concentrations are thought to be responsible for a significant part of the observed increase in average global temperature over the past 50 years (IPCC, 2007).

The total global soil carbon pool of 2000–3800Pg contains about 1550Pg of organic carbon and 950Pg of inorganic carbon. This soil C pool is about three times the size of the atmospheric pool (760Pg) and 4.5 times that of the biotic pool (500–650Pg) (Gruber et al., 2004; Janzen, 2004; Lal, 2004). According to Schlesinger (1997), 13–17% of the total soil carbon in tropical forests resides in the upper meter of soil and has a fast turnover time. Trumbore et al. (1996) suggested that the greatest losses of soil carbon due to climate change would be in tropical regions, where their measurements of radiocarbon content indicated the presence of a large pool of soil organic matter with a relatively rapid turnover time. Tropical rainforest fires and deforestation activities in the State of Mato Grosso, Brazil, have been releasing about 67 Tg C yr⁻¹ (DeFries et al., 2008).

Soil respiration is the primary process through which CO₂ fixed by land

¹The contents of this chapter have been published as F. B. Zanchi, M. J. Waterloo, B. Kruijt, J. Kesselmeier, F. J. Luizão, and A. J. Dolman,. Soil CO₂ efflux in Central Amazonia: Environmental and methodological effects. *Acta Amaz.* vol.42(2): 173–184, 2012. (doi:10.1590/S0044-5967201200020000)

plants is returned to the atmosphere. The soil CO₂ flux is derived from autotrophic and heterotrophic sources. Estimates of future CO₂ changes in the atmosphere depend largely on the feedback of terrestrial ecosystems to climate change, in particular on the balance of C uptake and subsequent loss from ecosystems in a warmer world (Trumbore, 2006). Trumbore (2006) also mentioned the discrepancy between the bulk soil respiration measurements conducted in tropical forest environments and process-based models that provide information about the separate contributions from autotrophic and heterotrophic sources.

The availability of such information is a key factor for understanding the response of terrestrial ecosystems to climate change, and it is crucial to understand the effects of variations in biophysical regulators of soil respiration to assess carbon balance in forest ecosystems. However, estimates of soil carbon are highly uncertain (Nakayama, 1990; Janssens et al., 2000) and difficult to measure because of differences between ecosystems and external drivers such as temperature Lloyd and Taylor (1994); Davidson et al. (1998); Janssens et al. (2003), soil moisture (Howard and Howard, 1993), soil texture and chemical properties (Trumbore et al., 1995; Liski and Westma, 1997), wind speed, leaf litter and root biomass (Reichstein et al., 2003; Trumbore, 2006) and activity of macro- and micro-fauna (Dantec et al., 1999; Giardina and Ryan, 2000; Raich et al., 2002). Several studies also suggest an influence of seasonal variation in litter fall on soil respiration rates (Reichstein et al., 2003; Salimon et al., 2004; Valentini et al., 2008).

On the other hand, in a study in the northwestern Amazon region, Metcalfe et al. (2007) found a weak correlation between soil efflux related to volumetric soil moisture ($R^2=0.44$) and non-correlation with temperature. There was a significant correlation among respiration from soil, litter, roots, and soil organic matter, so the heterotrophic and autotrophic contribution could be estimated.

Various methods have been used to study soil respiration in the Amazon region. Coutinho and Lamberti (1971), Martins and Matthes (1978) and Medina et al. (1980) measured the soil CO₂ efflux in upland terra firme rainforest and short statured Campina heath forest by capturing CO₂ in an alkali solution in a closed chamber on the forest floor. Wofsy et al. (1988) employed a headspace sampling technique and subsequent field CO₂ analysis of the air with a chromatograph to measure the soil CO₂ efflux. These techniques can be used to estimate long-term emissions, but are less appropriate for assessing variation in short time scales such as those typical for many biological processes. After 1990, soil respiration chambers were connected to infrared gas analyzers (*IRGA*) to form open- or closed-path and static or dynamic measurement systems (Fan et al., 1990; Chambers et al., 2004; Metcalfe et al., 2007). These modern systems allow assessment of the variation of soil respiration over shorter periods and provide reliable measurements of soil CO₂ efflux. Their operation and particularities have been described in various publications (Livingston and Hutchinson, 1995; Dantec et al., 1999; Davidson

et al., 2002; Pumpanen et al., 2004). It is often necessary to have an equilibrium period after ring insertion before starting soil respiration measurements (Hutchinson and Livingston, 1993; Livingston and Hutchinson, 1995), which was not always done (Chambers et al., 2004; Souza, 2004). Furthermore, Metcalfe et al. (2007), using a rhizotron to observe deep root growth, showed that in root-free soil it may take up to three months (after rhizotron installation) before root mass reaches a natural level and that the rate of root growth is approximately linear, whereas root mortality remains negligible.

The present study aimed to assess the influence of ring insertion and precipitation events on instantaneous measurements of soil CO₂ efflux. We also present an overview of soil respiration measurements in various Amazonian forest types and estimate the total soil CO₂ emission from the *Cuieiras* and *Campina* Reserves.

2.2 Material and methods

2.2.1 Experimental areas

The experimental areas were the Reserva Biológica do Cuieiras (Cuieiras Reserve), located at (2° 36' 32.67" S, 60°12'33.48" W, 110 m a.s.l. - above sea level), and in the *Reserva de Campina* (Campina Reserve), located at 2°35'30.26" S, 60°01'48.79" W, (65 m a.s.l.). These reserves are both under control of the National Amazon Research Institute (*Instituto Nacional de Pesquisa da Amazônia* - INPA).

The Cuieiras Reserve covers 22.7 hectares and is located about 70 km north of Manaus-AM, Brazil (Chambers et al., 2004; Araújo et al., 2002). The landscape consists of unconsolidated sedimentary layers which are dissected by rivers and streams, creating a pattern of rather flat plateaus (90–130 m a.s.l.) and swampy valleys (45–55 m a.s.l.), separated by moderately steep slopes (15–30°) (Waterloo et al., 2006) (see Figure 2.1a, for additional site details of the transect). The vegetation is a mosaic of evergreen forest with a canopy height of about 35–40 m, with emergent trees over 45 m tall, varying to ecotone Campinarana and valley or riparian forest, according to the slope. Such forests cover about 5–6% of Amazonia and depend on variations in soils, nutrients and drainage conditions (Luizão et al., 2007). The recent development of the Height Above the Nearest Drainage (*HAND*) descriptor based on the Shuttle Radar Topographic Mission (*SRTM*) elevation data allows the classification of terrain according to water table depth and topography citeprenno.ea:2008:hnt. Analyses of the *Igarape Asu* catchment (Cuieiras Reserve area) indicated that valley forest environments (riparian and Campinarana) cover 43% of the area, whereas slope and plateau (terra firme) forests occupy 26% and 31%, respectively. The Leaf Area index (*LAI*) measurements were performed with an LI-2000 apparatus (LI-COR, Nebraska, USA) during the wet and dry season in April, July–September 2007 and in the same

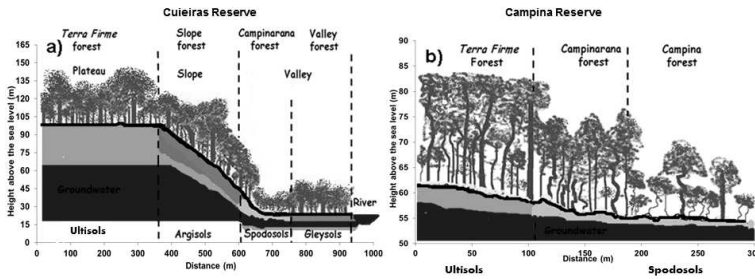


Figure 2.1: Typical topographic transect for tropical rainforest (Central Amazonia), where soil respiration was measured in different forest types. a) Cuieiras Reserve, which has a steep inclination forest and is composed of a mosaic of riparian, *Campinarana*, slope and plateau forests with their respective soil types; b) Campina Reserve, which presents the transition from plateau forest via *Campinarana* to *Campina* forest, with different in soil types and properties. After Chauvel et al. (1987).

periods of 2008. The measurement design consisted of two grids, one in the Campinarana and another in the riparian forest of Cuieiras and two others in the Campina Reserve. Both grids consisted of three parallel lines with 15 m spacing between them and 100 m in length. The measurements were made at one meter intervals and all values were averaged for the wet and dry season. The LAI of the plateau forest was 6.1 (Marques-Filho et al., 2005), that of the Campinarana forest was 5.0 ± 0.6 and that of riparian forest was 5.8 ± 0.6 .

Cuieiras Reserve has the following main vegetation types: the riparian area consists of *Arecaceae*, *Caesalpinaceae*, *Dichapetalaceae* and *Burseraceae* (I.L. Amaral, unpublished) families, which transitions to *Campinarana* forest, where *Caesalpinaceae*, *Euphorbiaceae* and *Sapotaceae* are dominant. In turn, the slope forest is dominated by *Lecythidaceae*, *Sapotaceae*, *Chrysobalanaceae* and *Burseraceae* and the plateau *terra firme* forest by *Lecythidaceae*, *Sapotaceae*, *Fabaceae* and *Euphorbiaceae* (Oliveira and Amaral, 2004, 2005). All these areas have their own singular characteristics and some *Campinarana* species occur in both *Campina* and dense *terra firme* forests (Proctor, 1999; Luizão et al., 2007). The transition from *Campinarana* to lowland evergreen rain forest is marked by the sudden appearance of palms in the understory and a drastic reduction of the root mat (Luizão, 1996), which is determined by the soil properties and different species composition.

Soils on the plateau are clayey *Oxisols*, whereas the slopes are dominated by *Ultisols*. Valley soils generally consist of strongly leached quartz sands, classified as *Spodosols* and *Gley* soils, with low water and nutrient retention capability (Brinkmann, 1985; Chauvel et al., 1987; Waterloo et al., 2006), high phenolic content and acidity (Proctor, 1999).

The Campina Reserve is located 60 km north of Manaus. The reserve cov-

ers approximately 900 ha, of which 6.5 ha is stunted heath forest (SHF) and tall heath forest (THF).

The SHF area (2.6 ha) is a mosaic of shrub islands surrounded by white bare sandy soil and 3.9 ha of the THF area, together with the rest of the reserve, consists of lowland evergreen rain forest. The canopy height is about 4–7 m for SHF, 10–18 m for THF and 25–35 m for *terra firme* forest (Luizão, 1996) (see, Figure 2.1b for additional site transect details). Using the Shuttle Radar Topographic Mission (SRTM) elevation data, we separated an area of 500 x 500 m and analyzed the georeferenced points. The image vegetation contrast was used to indicate the percentage of the vegetation area, which was 73% of the THF, 15% of the SHF, while 11% of the area was bare soil. The terra firme forest was not included in the analysis because it is located outside the selected area. The LAI was measured with an LI-2000 apparatus (LI-COR, Nebraska, USA) during April, July–September of 2007 and also in the same periods in 2008. The same method was used in the Campina Reserve as in the Cuieiras, but with only one grid in the SHF and another in THF. The SHF forest LAI was 2.4 ± 1.3 and for the THF it was 3.9 ± 1.0 .

The main characteristics of the Campina Reserve are the structure and scleromorphic physiognomy that distinguish heath forests from other regional forests types. These forests grow on extremely nutrient-poor white sandy soils (spodosols) and exhibit low species richness, dominated by one or more species, and an unusual physiognomy: shorter stature, many branched and tortuous trees and bushes with scleromorphic leaves and considerable load of vascular epiphytes (Orchidaceae, Bromeliaceae, Araceae, Ericaceae) and lichens (Anderson, 1981; Richards, 1996). The shrub and tree species are *Ouratea spruceana* (Ochnaceae), *Pagamea duckei* (Rubiaceae), *Pradosia schomburgkiana* (Sapotaceae), *Adina heterophylla* (Caesalpinaceae) (Anderson, 1981; Luizão, 1996).

But there were also species which belong only to these areas due to adaptation to soil type, micro climate and other specific environmental characteristics. For example, the Campina Reserve shares only 3.2% of species with *terra firme* forest and 17% with the Campinarana forest (Oliveira and Amaral, 2004).

The spatial and temporal variation of climate in the central Amazon region is low. The annual average temperature is 26.7 ° C, with relative humidity at about 80%. The annual rainfall (1966–1992) reported about 75 km SE from the Campina Reserve (Ducke Rainforest Reserve) amounts to 2442 mm, with a standard deviation of 306 mm (Waterloo et al., 2006). The dry season occurs from June to November. The Campina Reserve is warmer than the Cuieiras Reserve, because the forest is more open, contains a less dense canopy and is composed of small islands of bushes, allowing more wind and heat to enter the understory.

2.2.2 Soil measurements

Soil respiration was measured randomly using an LI-8100 automated soil CO₂ flux system (LI-COR, Nebraska, USA) attached to an LI-8100-101 long-term chamber (20 cm in diameter). Measurements were also performed using another automated system developed by Alterra Institute, The Netherlands. This system consists of two chambers measuring 25 cm high and 30 cm in diameter made of polyethylene coupled to a CO₂ an LI-840 analyzer (LI-COR, Nebraska, USA). Both systems have output in $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$.

The measurement methods were based on dynamic chambers (Norman et al., 1992) and measurements were taken once every 30 minutes during 24 hours. Following (Hutchinson and Livingston, 1993), the systems were installed at 2 cm depth into clayey soils and 3 cm into sandy soils.

Measurements using the LI-8100 system were made in one place chosen randomly during a week and then the system was moved to another place inside of the measurement area or to a different location inside of the Cuieiras Reserve or the Campina Reserve (Figure 2.1). The measurements in the Campina area were carried out between 20 July and 11 October 2007. The measurements in the Cuieiras Reserve were performed as follows: on 8 February 2007 and between 13 November and 13 December 2007 in the riparian area; between 20 November 2006 and 31 May 2007 in the Campinarana area; between 28 June and 12 July and between 12 and 22 December 2006 in the slope area; and between 3 August and 6 November 2006 and again between 21 and 26 February 2008 in the plateau area.

Measurements were made with a second soil respiration system (the Alterra system) in the Campinarana forest of the Cuieiras Reserve. This system consists of a chamber with a lid closed by means of an electric motor with pulley and rope. The measurements are taken when the lid closes automatically at programmed measurement intervals. Inside the chamber the system has a fan to mix the air (Micronel, D341T, Vista, USA). The air enters the chamber through polyethylene tubing placed at the base and leaves through a similar tube below the chamber lid. There is also a pressure balancing tube which equalizes the pressure between the chamber and the outside air. The lid is sealed with two hollow neoprene bands to prevent any external contamination.

The LI-8100 system measurements were performed one week after the ring installation in order to prevent any error in the results that could be attributed to the soil disturbance. On some occasions spare polyvinyl chloride (PVC) rings were not available to be installed before the random measurements. Soil ring insertion produced outliers, which were removed by statistical analyses for the number of standard deviations from the mean. However, to identify the ring insertion effects we used the first data set compared with the next three days of measurements and the outliers were calculated by the increase compared to the average results.

Effect of rainfall events on soil respiration was determined using data obtained 24 hours before the event for comparison with those measured during

24 hours after the events. These data were compared for different amounts of precipitation input and for all sites measured.

For the distribution of the spatial variation of soil CO₂ respiration, we used Shuttle Radar Topographic Mission (SRTM) images. The areas covered by the different forest types were determined, and combined with local measurements of soil CO₂ respiration, to estimate the total area-averaged emission for each reserve.

Near the soil chambers, soil moisture and temperature sensors were installed at 10 cm depth and 5 and 10 cm depths, respectively. Both sensors measured at the same frequency as the LI-8100 and the Alterra systems.

2.2.3 Statistical analysis

To verify normal distribution on the data, the *Kruskal-Wallis* test and one-way analysis of variance (ANOVA) by ranks were applied. This is a non-parametric method for testing the equality of population medians among groups of soil respiration from all different measured areas. Because it is a non-parametric method, the test does not assume a normal population. However, it does require an identically shaped and scaled distribution for each group, except for any difference in medians. This test was used because another statistical test was performed in relation to the mean and the possible outlier data from rain events or some high or low efflux emission measurements that could reduce the statistical significance of the analysis.

Regression analysis was performed to examine the relationship between soil respiration and soil moisture and to ensure that the residuals were equally distributed about the regression line with constant variance (homoscedasticity) where the significant effects were determined ($p < 0.05$). All statistical analyses were performed using the Matlab software (version 7.0, The MathWorks, Inc., USA).

2.3 Results and Discussion

2.3.1 Ring insertion effect

Determining the variability of soil CO₂ efflux requires use of an appropriate measurement method and measurements need to be taken in many different locations (Sotta et al., 2004). The final results can be underestimated if the methodology used is not appropriate. This study found that ring insertion caused an increase of about 13–20% in the mean value of soil CO₂ efflux for the first seven hours (Figure 2.2). Keller et al. (2000) reported that ring installation might damage the soil by breaking the micro- and macro-soil chambers, including roots, and can release high concentrations of CO₂, NO, N₂O and CH₄. However, soil has some mechanisms that can minimize these emission

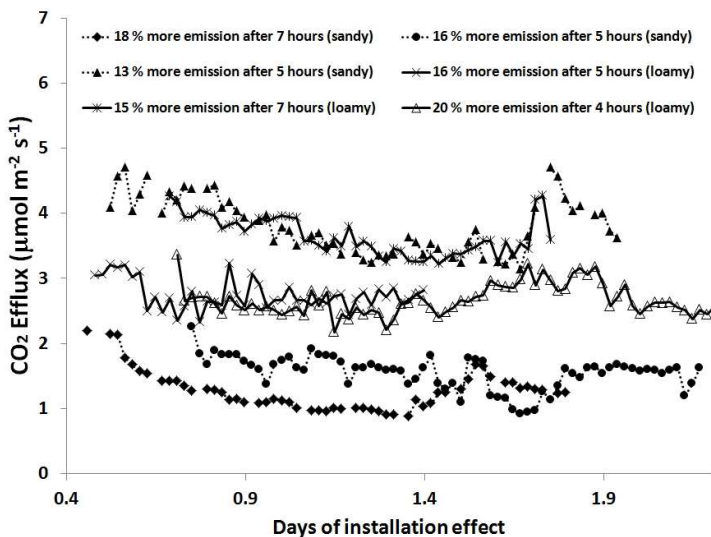


Figure 2.2: Effect on the soil CO₂ emission after ring installation (soil insertion) for all different measurement sites in the *Cuieiras* and *Campina* Reserves. The time-axis is in fractional days after ring installation following the start of CO₂ measurements.

effects and reduce the time scale of the artificial influence, (e.g., photosynthesis, transpiration, precipitation, physiological processes and decomposition rates). Metcalfe et al. (2008) reported a high root increase over a short period. In particular fine roots, which can grow more than 2 cm per day, are an important indicator that soil can recover quickly (after about seven hours) in tropical surface layers (2 cm depth). Nevertheless, it takes some time until the natural soil CO₂ emissions are established again, depending on soil and forest types. Across the surveyed sites, these artificial mechanical factors caused extra emissions and the majority of emission variation persisted between four to seven hours after ring installation. The loamy soils ($n=5$, 3 in the plateau and 2 in the slope forest, $p<0.05$) showed an increase of 15–20%. The sandy soil ($n=6$, 4 in the *Campina* and 2 in riparian Cuieiras Reserve area, $p<0.05$) also showed an increase of 13–18%.

These variations in CO₂ emission are not important if the measurements are performed over long periods. This can be achieved by using an automatic system for 24 hours. We noted that a maximum of seven hours after the insertion, the effluxes stabilized to normal emission rates for all soils measured at the study sites. On other hand, when the system was used randomly and the measurement is taken punctually after ring insertion, the overestimation can add a few tonnes to the annual carbon emission estimates (Chambers et al., 2004; Souza, 2004). In general, the length of the waiting period is uncertain

(Davidson et al., 2002) and it is necessary first to check how long the ring insertion effect is likely to influence the mean results.

2.3.2 Precipitation effect

There was a considerable variation in the soil CO₂ efflux related to precipitation events. The efflux for all soils and forest types also behaved differently.

Nevertheless, there were no large changes observed in plateau and slope areas (clayey soils, rainfall below 2 mm, n=4), but there was an increase of less than 10% in the first hour after the precipitation event. Then the soil CO₂ efflux returned to normal. Sotta et al. (2004) reported that soil respiration responded somewhat to precipitation events of less than 2 mm. In this study, we also found an increase in the soil CO₂ efflux related to rainfall, but not for precipitation below 2 mm (F. B. Zanchi, in prep., interception data from 2006–2007). We noticed that this amount of rainfall generally did not reach the forest ground, and therefore it could not cause any change in the soil respiration, temperature and moisture in the top layer.

When the rainfall was between 3 to 8 mm (n=3), we noted an increase in the emission of about 10 to 18% for the next 2 hours. After this period, the efflux dropped by 15% compared to the normal efflux and returned to normal after 17 hours. During this period the soil temperature changed slightly.

With rain events greater than 8 mm (Fig. 2.3), we noticed two large changes in the soil emission. The total of the first event was about 12 mm and the soil moisture in the top 10 cm increased from 0.30 to 0.39 m³ m⁻³ in 2 hours, and there was a small spike in the soil CO₂ in the same period after the rainfall event. After 2 hours, the soil moisture started to decrease and the soil respiration decreased due to the groundwater percolation through the soil pores. The soil CO₂ emissions only returned to a normal diurnal cycle after 18 hours (about 22% less), caused by the rainfall effect. Likewise, after 22 mm of rainfall, we noticed that the emission was suppressed immediately after the event, by about 27%, followed by a decrease in soil temperature (Fig. 2.3).

On the other hand, we found different variations for sandy soils. In the Campinarana forest (Cuieiras Reserve, Fig. 2.4), a rainfall event smaller than 3 mm did not change the CO₂ emission. Between 3 to 10 mm rainfall (n=11), the respiration increased quickly by 34% during 2 hours and returned to normal after 4 hours. Rainfall greater than 15 mm (n=2) caused a quick decrease in CO₂ efflux, an effect that lasted for 5 hours at low rates, while the soil moisture also changed quickly from 0.1 to 0.15 m³ m⁻³ during the same period. The soil temperature showed a slight decrease in this period (Fig. 2.4).

For the same sandy soils but in a different forest composition (Campina Reserve), the *THF* and *SHF* soil respiration behaved similarly, showing little change compared to the mean emission (<10% for 15 hours). Fig. 2.5 and 6 show a slight decrease followed by a large spike for 2 hours after the rain event, which decreased the amplitude of the soil CO₂ diurnal cycle. The soil temperature also decreased and there was a quick increase for a few hours in

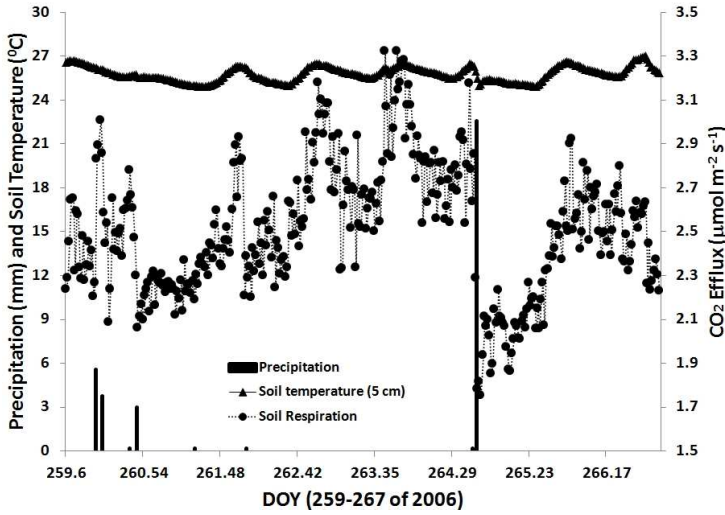


Figure 2.3: Change in the soil CO₂ efflux emission for the loamy soil of the plateau forest from Cuieiras Reserve caused by precipitation. The effect decreased the emission by 25% (n=4) for 18h after the rainfall event compared with the previous emission. In days of year (DOY) from 259-267 of 2006.

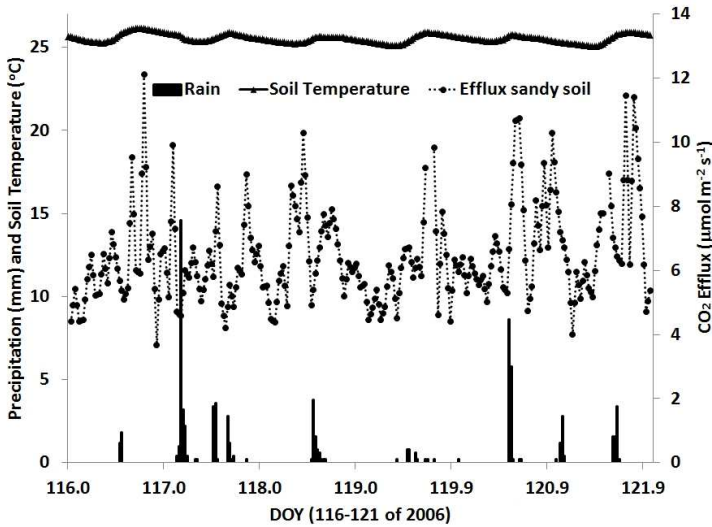


Figure 2.4: Effect of rainfall events on soil CO₂ efflux emission for the sandy soil of *Campiranara* forest from *Cuieiras* Reserve. The effect decreased the emission by 7% (n=6) in relation to the previous emission around 7h after the rainfall events. In days of year (DOY) from 116–121 of 2006.

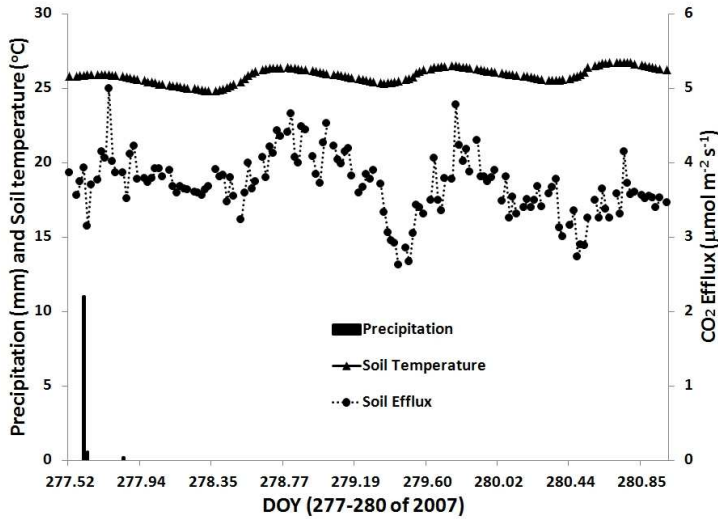


Figure 2.5: Rainfall influence on the soil CO₂ efflux emission. The rainfall did not change the mean CO₂ emission for the sandy soil of *THF* forest in *Campina* Reserve, but the efflux had some effect about 2h after the rainfall event, with an abrupt increase before returning to the regular emission level again. In days of year (DOY) from 277-280 of 2007.

the soil moisture, perhaps because of the sandy soil and small CO₂ production in this type of forest.

Two hours after big rainfall events, soil CO₂ emission decreased by 10 to 27%. Similar results were found by Matteucci et al. (2000), Sotta et al. (2004) and Savage et al. (2009), all of whom found this rainfall influence, but for other soil types.

In addition, there were abrupt changes in the soil temperature when the rainfall event was bigger than 10 mm (Fig. 2.3 to Fig. 2.6). These soil CO₂ efflux increases were noticed in all 7 analyzed events. Perhaps the reason of this increase is more a physical effect than an effect of groundwater percolation derived from 2 mm rainfall, because the air pressure dropped around 3 to 5 mbar just before and during the rainfall events. We could not find any other strong relation for small precipitation to ascertain which factor affected the soil evasion other than atmospheric pressure. And the amount of rain was not big enough to provoke such high spikes and influence the soil evasion. Some authors, such as Hutchinson and Livingston (1993), Lund et al. (1999) and Davidson et al. (2002), have stated that slight changes in atmospheric pressure can influence soil CO₂ emissions. Maybe this was the reason small rainfall amounts caused a change in the soil CO₂ emissions in this study.

However, it is still unclear if the precipitation was the main factor influenc-

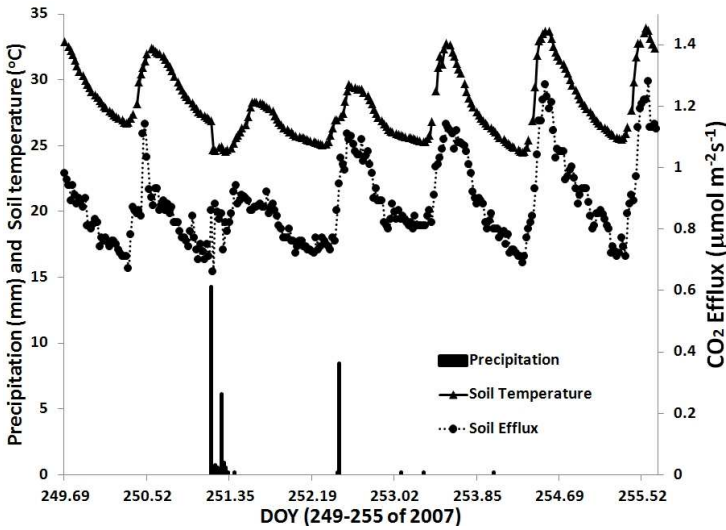


Figure 2.6: Effect of rainfall events on the soil CO₂ efflux emission. The precipitation did not change the mean CO₂ emission, but the emission had a slight effect on the soil respiration, followed by soil temperature decrease for the SHF forest from *Campina* Reserve. In days of year (DOY) from 249-255 of 2007.

ing the soil CO₂ emission due to high/low soil moisture content. But we can also hypothesize that the bacterial decomposition of organic matter and the water filling of soil pores may trigger an increase in CO₂ emission, as mentioned by Linn and Doran (1984). Also, atmospheric pressure also can have an important influence on the CO₂ emissions Hutchinson and Livingston (1993).

2.3.3 Spatial distribution

There was a significant variation between sites in the recorded CO₂ efflux variables (Table 2.1). For 23 different Amazonian sites, the mean value was $4.2 \pm 1.8 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ and the correlation coefficient was greater than 42% among all the sites. Table 2.1 and Fig. 2.7 show an example from two small catchments, where we found that the soil CO₂ variation between soil types and vegetation was high. Inside the *Campina* reserve, which contains bare soil along with SHF and THF areas, the variations were 0.99 ± 0.14 , 1.2 ± 0.3 and 3.8 ± 0.2 respectively. In the *Cuieiras* Reserve, where a transect from the upper to lower area of around 900 m in length is present (Fig. 2.1), we also found the highest differences between the vegetation and the soil types. Consequently, the soil emission rates showed large differences, which changed in relation to plateau, slope, *Campinarana* and riparian forests and where the soils vary from clayey to sandy. In these areas, the soil respiration levels were

2.4±0.4, 3.2±0.5, 6.0±1.6 and 4.1±1.2 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$, respectively.

The Campinarana and the THF sites inside each reserve had the highest emissions. Both microenvironmental areas have similar vegetation and soil type. The main characteristics of these sites differ from other locations in this study, by having a thick root matter and litter layer in the first 5 to 10 cm of soil. The emissions were higher here than at all the other sites.

To quantify the total CO_2 efflux from the Campina and Cuieiras reserves, we calculate the weighted average according to the size of each vegetation type. Assuming that the vegetation of the reserves grows in each soil type, we described the areas using the SRTM images. The images provided a suitable mesh for each forest size from the Campina Reserve, where THF= 73%, SHF= 15% and bare soil= 11% of the total area. The weighted average of the whole Campina Reserve was found using the equation $R_tCP = 3.83 * 0.73 + 1.2 * 0.15 + 0.99 * 0.11$, and the total Campina emission was $3.08 \pm 0.8 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$. The Cuieiras Reserve's weighted average was estimated using the same method, plateau = 31%, slope = 26% and riverine (Campinarana and riparian forest) = 43% of the area. The equation was $R_tCU = (2.4 * 0.31) + (3.2 * 0.29) + (5 * 0.43)$, with a total emission of $3.82 \pm 0.76 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ for the entire area of Cuieiras Reserve.

An important improvement compared to previous studies of the Cuieiras Reserve (Chambers et al., 2004; Sotta et al., 2004; Souza, 2004) and Campina Reserve (Martins and Matthes, 1978) was the fact we managed to quantify the CO_2 emissions in relation to the vegetation and soil types.

The spatial variation results for all the measured sites (Fig. 2.7, $F = 18479$ *d.f.* = 5, $p = 0$) showed a strong indication that the spatial variations for testing equality of population medians among groups were not the same, with a difference prevailing between all the sites measured.

To find the relation of soil temperature or the main driven factor on the soil respiration, we selected from the dataset only periods without rainfall. An Exponential model adjusted for a daily cycle of soil respiration with soil temperature dependence at 5 cm. We noted a hysteresis on both variables. Soil respiration answer faster to the physical parameters compared to the soil temperature. To obtain a better fit curve, was necessary to shift-back around 3 hours the soil temperature related to soil respiration, which could predict the time hysteresis (Gaumont-Guay et al., 2006; Pavelka et al., 2007; Pingintha et al., 2010a). The simple model turns to, $Rs = 0.0598 * e^{0.1435 * Temp}$, with the $R^2 = 0.9$ ($p < 0.05$). Afterwards the model was applied for the whole soil temperature dataset, showing a similar pattern for both and a better fit curve.

Sotta et al. (2004) also found that rainfall events explained 75% of the correlation between soil CO_2 and soil temperature decreases. We assumed that the main driver of the soil CO_2 efflux for all the sites was soil temperature. While rainfall had an influence in both decreases, the main factor influencing emission was soil temperature.

Table 2.1: Soil CO₂ efflux measurements using different methods in Amazonian sites. Soil CO₂ efflux in $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ and soil temperature in °C. (Sotta et al. (2004) Adaptation).

Author	Season	Location	Vegetation	Soil Temperature	CO ₂ Efflux	Methodology
Coutinho and Lamberti (1971)	Dry (Aug–Sep)	Barcelos, AM, Brazil	Closed Rainforest	25–28	2.8	0.5N KOH solution
Martins and Matthes (1978)	Dry (Jul)	Manaus, AM, Brazil	Campinarana/Campina	Not stated	1.4±0.5	0.5N KOH solution
Medina et al. (1980)	2 years long	San Carlos do Rio Negro, Venezuela	Laterite Forest	22–27 forest floor	3.1±0.5	0.5N KOH solution
Wofsy et al. (1988)	Dry (Jul-Aug)	Res. Ducke, Manaus-AM, Brazil	Closed Rainforest	Not stated	4.5	Chromatography-CSS
Fan et al. (1990)	Wet (Apr–May)	Res. Ducke, Manaus-AM, Brazil	Closed Rainforest	Not stated	5.9	IRGA-CDS
Kepler et al. (1990.)	Not stated	Fucada, Manaus-AM, Brazil	Forest, Pasture and burned forest	24	6,5± 0.6	IRGA-CSS
Meir et al. (1996)	Wet-dry (May-Jun)	Res. do Jarú-RO, Brazil	Open Rainforest	22.9	5.5±1.6	IRGA-CDS
Trumbore et al. (1995)	Year long	Paragominas-PA, Brazil	Open Rainforest		6,1	IRGA-CDS
Davidson et al. (2000)	Year long	Faz. Vitória, Paragominas-PA, Brazil	Open Rainforest	22–24 (10 cm)	5.3	IRGA-CDS
Metcalfe et al. (2007)	Year long	Caxiuanã-PA, Brazil	Closed Rainforest	25±5 (5 cm)	3.6	IRGA-CDS
Chambers et al. (2004)	Year long	Manaus-AM, Brazil	Open Rainforest	Not stated	3.8	IRGA-CSS
Souza (2004)	Year long	Manaus-AM	Open Rainforest	24.5 (5 cm)	5,76	IRGA-CDS
Sotta et al. (2004)	End of wet season	Manaus-AM, Brazil	Closed Rainforest	25.6 (5 cm)	6.4±0.2	IRGA-ODS
Salimon et al. (2004)	Wet and dry	Rio Branco-AC, Brazil	Closed Rainforest and pasture	23.8±0.8 (5 cm)	4.73	IRGA-CDS
Valentini et al. (2008)	Year long	Sinop-MT, Brazil	Open Rainforest	24 (5 cm)	7.6±0.5	IRGA-CDS
Goulden et al. (2004)	Year long	Tapajos-PA, Brazil	Closed Rainforest	26 (5 cm)	3.38	IRGA-CSS
Present study	Wet and dry	<i>Cuiéiras</i> Reserve, Manaus-AM, Brazil	Closed Rainforest	25.7±0.6 (5 cm)	2.4±0.4	IRGA-CDS
			Slope Forest	25.3±0.3 (5 cm)	3.2±0.5	
			<i>Campinarana</i>	25.6±0.5 (5 cm)	6.0±1.6	
	Wet and dry	<i>Campina</i> Reserve, Manaus-AM, Brazil	Valley Forest	25.8±0.6 (5 cm)	4.1±1.2	
	<i>Campina</i> THF		25.8±0.4 (5 cm)	3.8±0.2		
	<i>Campina</i> SHF		26.7±2.6 (5 cm)	1.2±0.3		
<i>Campina</i> Bare soil	28.4±2.2 (5 cm)	0.9±0.1				

CDS = closed dynamic system; CSS = closed Static system; ODS = Open dynamic system
 ±x.x = standard deviation

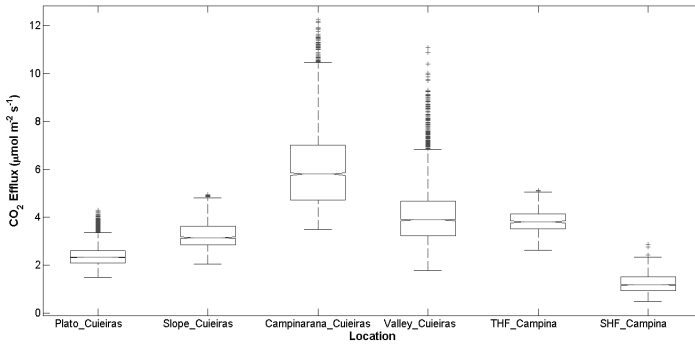


Figure 2.7: Topographical gradients from the rainforest in the central Amazonia along which CO₂ were measured in different locations of the *Cueiras* and the *Campina* reserves. The boxplot describes the minimum sample, lower quartile, median, upper quartile and maximum emission sampled from each location.

2.4 Conclusions

To quantify the total CO₂ emission, were important to have all information about the underlying controls upon respiration from different soil and vegetation (e.g. soil disturbs and rainfall intensity). On the other hand, by not applying the correct methodology, this study indicated an overestimate of 20% in the total soil CO₂ emission.

To minimize the soil disturbance and waiting time after ring installation before starting measurement, it is essential to know the soil type and vegetation characteristics. Comparing the loamy with sandy soil, in general the loamy soil was more sensitive to mechanical disturbance and took longer to return the natural emission level.

Rainfall events showed different patterns. The loamy soils were more stable than the sandy soils. For moderate rainfall events (8 mm), the increase in CO₂ efflux from sandy soil was 50% faster than for the loamy soils, and decrease after 2 to 3 hours, which did not happened in the loamy soils.

On the other hand, we found that the main factor influencing the soil respiration was soil temperature, because the soil respiration followed the same pattern as the temperature, while rainfall only caused a brief disturbance in the soil respiration. Rainfall seems to contribute only by creating favorable conditions for a quick decrease in temperature and consequently the respiration followed the physical effect of soil water percolation. We also noticed that it was difficult to account for the rainfall effects in the simple model estimation. This has some implications for future modeling studies over space and time, because rainfall causes overestimation of the total model emission and it is not easy to include the CO₂ emission spikes in the models.

Finally, the quantification of the total increase in CO₂ emission was better estimated using the total weighted average area compared to the local measurements from previous estimation studies. Thus, for better estimation of soil CO₂ effluxes, or to model a region and vegetation type, it is necessary to find the main influencing factors to decrease the uncertainties about the final carbon release measurements.

Chapter 3

Soil CO₂ exchange in seven pristine Amazonian rain forest sites in relation to soil temperature²

3.1 Introduction

Soil respiration (soil CO₂ efflux or R_s) is the sum of autotrophic (living roots) and heterotrophic respiration (Metcalf et al., 2011). Heterotrophic respiration, the microbial respiration of organic matter or mycorrhizae and some microbes subsisting on root exudates is the final phase of the mineralization of organic substances and carbon turnover. It can be regarded as a direct proxy of the energy passing through biota (Kuziyakov and Gavrichkova, 2010). Knowledge of the carbon turnover is important to understand the fate of the soil carbon pool (C). This carbon pool is estimated to be globally about three times the size of the atmospheric pool (760 Pg) and 4.5 times the biotic pool (500–650 Pg) (Gruber et al., 2004; Janzen, 2004; Lal, 2004). The global CO₂ emission is the strongest terrestrial carbon source to the atmosphere (Raich et al., 2002), about $89 \pm 12 \text{ Pg C y}^{-1}$, increasing by 0.1 Pg C y^{-1} related to the rising air temperatures (Raich et al., 2002; Bond-Lamberty and Thomson., 2010).

The contribution of the Amazonian rainforest to the carbon cycle is of special importance because of its huge primary production (18 Pg C y^{-1}) (Malhi

²The contents of this chapter have been submitted as F. B. Zanchi, A. Meesters, M. J. Waterloo, B. Kruijt, J. Kesselmeier, F. J. Luizão and A. J. Dolman. Analytical modeling of soil CO₂ exchange within the Central Amazonian rain forest in relation to soil temperature, submitted to Agriculture and Forest Meteorology

and Grace, 2000). The Amazonian rainforest carbon cycle has been investigated with many different techniques, such as eddy-covariance measurements (Goulden et al., 2004; Araújo et al., 2010), soil chamber measurements (Meir et al., 1996; Sotta et al., 2004; Salimon et al., 2004; Zanchi et al., 2011), biomass assessments (Malhi et al., 2009a), hydrological carbon export studies (Richey et al., 2002; Waterloo et al., 2006), remote sensing and airborne flux studies (Lloyd et al., 2007). However, upscaling the results is difficult (Malhi et al., 2009a; Araújo et al., 2010; Davidson et al., 2012), because the Amazonian area is very large (about 6.6 million km²), and heterogeneous in vegetation, soil type and micro climate (Phillips et al., 1998; Zanchi et al., 2012).

There is a broad consensus about the need to validate models for ecosystems, especially for the tropical region due to climate anomalies and ecosystems with significant total carbon stocks (Nobre et al., 1996; Waterloo et al., 2006; Malhi et al., 2009b). As only local measurements are available, one has to rely on modeling, employing the relations which can be found for specific ecosystems. Thus, measurements are essential to get a reliable understanding of the forces driving soil respiration. This may help to further reduce the uncertainties in the sensitivities of various models (Davidson et al., 2002; Zanchi et al., 2012).

However, the relations with the control variables (abiotic parameters) in the Amazonian R_s are not yet sufficiently understood (Metcalf et al., 2007) or even measured (Zanchi et al., 2012). In general, soil respiration has been shown to be controlled not only by soil temperature (T), but also by other environmental variables such as soil moisture, soil texture and composition, chemical properties and biomass of leaf litter and roots, tree size, as well as the activity of macro and microfauna (Howard and Howard, 1993; Liski and Westma, 1997; Davidson et al., 1998; Irvine and Law, 2002; Bréchet et al., 2011). A major problem here is that root respiration and root exudate supply are driven by photosynthesis, which has only a remote connection to the soil temperature (Tang et al., 2005; Baldocchi et al., 2006; Gaumont-Guay et al., 2008).

Though Amazonian studies have tried to find how variations of R_s are controlled by biotic and abiotic variables (*e.g.*, Sotta et al. (2004); Salimon et al. (2004); Metcalf et al. (2007); Bréchet et al. (2011); Zanchi et al. (2012)), the relationships have appeared hard to disentangle. The main reason is that R_s measurements made over longer periods of time with seasonal information are still scarce in the Amazon region (Zanchi et al., 2012).

Soil temperature is one of the most important physical variables that control soil respiration. It is usually conveyed in the Q_{10} factor (soil respiration rate changes when temperature increases by 10°C) (Lloyd and Taylor, 1994; Reichstein et al., 2003). Use of the Q_{10} -factor to model the soil temperature dependence is somewhat involved, because the ecosystem dynamics can also cause time lags (hysteresis) (Baldocchi et al., 2006; Vargas and Allen, 2008) (Phillips et al., 2011). These time lags (different phase between soil respiration

and soil temperature) have been already indicated by several authors (Pavelka et al., 2007; Graf et al., 2008; Subke and Bahn, 2010; Zanchi et al., 2012). The mechanisms that regulate this time lag might be driven by rhizosphere dynamics (Gaumont-Guay et al., 2008; Vargas and Allen, 2008), plant phenology and photosynthesis (Tang et al., 2005; Baldocchi et al., 2006; Vargas et al., 2010). A major difficulty for determining the key controlling factors (Kuzyakov and Gavrichkova, 2010) is the lack of accuracy in determining the seasonal variations in the biophysical lags (Högberg et al., 2001; Baldocchi et al., 2006; Graf et al., 2008; Vargas and Allen, 2008).

To minimize this ecosystem hysteresis, some authors have used empirical observation methods or gas diffusion time lag measurements (Vargas and Allen, 2008; Pingingtha et al., 2010a) or analytical solutions (Gao et al., 2008; Graf et al., 2008; Sierra et al., 2011), Fourier or alternative tools such as wavelet or stepwise analysis (Tang et al., 2005; Graf et al., 2008; Vargas et al., 2010) to analyze the dependencies between the data sets.

A further methodological problem related to the time lag is that it depends on the choice of the soil temperature measurement depth (Graf et al., 2008; Vargas et al., 2010). Assuming an arbitrary measurement depth of soil temperature, yields an apparent Q_{10} value in the soil temperature sensitivity (Phillips et al., 2011), because of the phase shift and also the amplitude decrease of the temperature (Graf et al., 2008). However, most studies use a single and somewhat arbitrary temperature measurement depth, and the effect of varying temperatures with soil depth is often not considered (Meir et al., 1996; Salimon et al., 2004; Sotta et al., 2004; Pinto-Junior et al., 2009).

Gaumont-Guay et al. (2006) suggest that the $T-R_s$ curve with the lowest hysteresis indicates the most appropriate temperature measurement. Pavelka et al. (2007); Zanchi et al. (2012) also found some hysteresis and recommended to find the maximum R^2 correlation, by performing a cross-correlation analysis to align the phase of soil depths temperature time series with R_s . This cross-correlation analysis is valid for a short period of data (no seasonality), but application to a series of a whole year is problematic because changes in soil thermal and CO_2 diffusivity, can cause a shift in the CO_2 efflux activity in the soil profile (Graf et al., 2008; Subke and Bahn, 2010).

The main aims of this study are to identify how the respiration for seven pristine tropical rainforest areas in Central Amazonia depends on the site characteristics and on the daily course of the temperature. Averages are considered for the wet and dry season separately. Further, the short-term relationship between soil temperature and R_s is quantified, accounting for the diurnal time lag. To perform this, a simple analytical treatment of the soil temperature is used, which can also be applied for other regions.

3.2 Material and methods

3.2.1 Site description

The study was executed in the experimental areas *Reserva Biológica do Cuieiras-ZF2* (2° 36' 32.67 S, 60° 12' 33.48 W, 40–110 m, a.s.l.—above sea level) and the *Reserva Biológica de Campina* (2° 35' 30.26 S, 60° 01' 48.79 W, 93–101 m, a.s.l.). Both are managed by the *Instituto Nacional de Pesquisas da Amazônia*–INPA (Table 3.1).

The Cuieiras Reserve covers 22,735 ha and is located about 70 km north of Manaus–AM, Brazil (Araújo et al., 2002; Chambers et al., 2004; Zanchi et al., 2011, 2012). The landscape consists of unconsolidated sedimentary layers which are dissected by rivers and streams, creating a pattern of rather flat plateaus (90–130 m a.s.l.) and swampy valleys (45–55 m a.s.l.), separated by moderately (15°–30°) steep slopes (Waterloo et al., 2006). The vegetation consists of a mosaic of evergreen forest with a canopy height of 35–40 m, with emergent trees over 45 m tall (Prance, 1979), varying from plateau and slope forests to ecotone of the *Campinarana* and valley or riparian forest (Zanchi et al., 2012). The *Campinarana* forest covers about 5–6% of the total Amazon region and grows on soils with thick root mat layer, variable nutrient and drainage conditions (Luizão et al., 2007). In the Cuieiras Reserve the vegetation consists in the valley area of *Arecaceae*, *Caesalpinaceae*, *Dichapetalaceae* and *Bursaceae* families (I. L. Amaral, unpublished), followed by the *Campinarana*, which contains *Caesalpinaceae*, *Euphorbiaceae* and *Sapotaceae*. The slope forest contains *Lecythidaceae*, *Sapotaceae*, *Chrysobalanaceae* and *Bursaceae*, while the plateau (*terra-firme*) contains *Lecythidaceae*, *Sapotaceae*, *Fabaceae* and *Euphorbiaceae* (Oliveira and Amaral, 2004, 2005) (Fig. 3.1, a)).

Rennó et al. (2008) using satellite image from Cuieiras Reserve, found that valley forest environments (Riparian and ecotone *Campinarana*) cover 43% of the Cuieiras reserve, whereas slope and plateau (*terra-firme*) forests occupy 26% and 31%, respectively. Soils on the plateau are clayey Oxisols, whereas the slopes are dominated by Ultisols. The clay-content of the soil surface (0–5 cm) decreases from about 75% to 5% with the sand-content increasing from 10% to 85%, when moving from plateau to valley (Ferraz et al., 1998). Valley soils generally consist of strongly leached quartz sands, classified as Spodosols and Gleys, with low water and nutrient retention capacity (Brinkmann, 1985; Chauvel et al., 1987; Waterloo et al., 2006), and high phenolic content and acidity (Proctor, 1999).

The *Campina* Reserve is located 60 km north of Manaus comprising approximately 900 ha. The forest types consist of stunted heath forest (SHF), formed by a mosaic of shrub islands surrounded by white sandy bare soil, tall heath forest (THF) and by a lowland evergreen rain forest (*terra firme*). The canopy height is about 4–7 m for SHF, 10–18 m for THF and 25–35 m for *terra-firme* forest (Luizão et al., 2007). Zanchi et al. (2012) separated in this Reserve a catchment area of 25 ha and analyzed the georeferenced points,

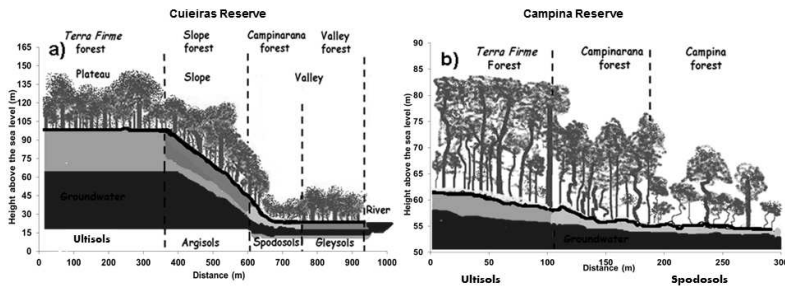


Figure 3.1: Typical topographic transect from tropical rainforest from Central Amazonia: a) Cuieiras Reserve, composed of a mosaic of valley, Campinarana, slope and plateau forests with their respective soil types; b) Campina Reserve, composed of a mosaic of plateau forest, Campinarana and Campina forest, and their differences in soil types and properties. Redrawn from Zanchi et al. (2012).

where the vegetation contrast through satellite image was used to indicate the percentage of the vegetation area. These authors found 15% of SHF, 11% of bare soil and 73% of the THF and *terra-firme* forest, respectively.

The *Campina* reserve has the structure and scleromorphic physiology that distinguishes heath forests from the other regional forests types. They are about 5–6% of the total Amazon region and have extremely nutrient-poor white sandy soils (Spodosols) with a low species richness and an unusual physiognomy dominated by short stature trees and bushes which host a variety of vascular epiphytes and lichens (Anderson, 1981; Richards, 1996). The shrub and tree species consist of *Ouratea spruceana* (Ochnaceae), *Pagamea duckei* (Rubiaceae), *Pradosia schomburgkiana* (Sapotaceae), *Adina heterophylla* (Caesalpiniaceae) (Anderson, 1981; Luizão, 1996). Luizão et al. (2007); Zanchi et al. (2012) also described the main characteristics of the *Campina* Reserve forests and the soil classifications (Spodosols), including species which belong only to these particular areas due to specific adaptations to the soil, micro climate and specific environmental characteristics. Luizão et al. (2007), showed that soils in this area are predominantly sand (with a high drainage status), varying from 90–99% of sand with some clay content of 3.6–0.4% in the first 1 meter of the soil profile and with very little soil carbon content in the first 10 cm, varying from 31% in the THF and 0.2% in the SHF (Fig. 3.1, b). According to Luizão et al. (2007); Zanchi et al. (2012) *Campina* Reserve can be drier than the Cuieiras Reserve, because the forest is more open, giving rise to higher air temperature. It also has a less homogeneous canopy cover, formed by small islands of bushes, allowing more wind and heat to enter the understory which increases the evaporation of water from the soil and vegetation.

The dry season occurs from June to September for both reserves. The

spatial and temporal climate in the central part of the Amazon region does not change much (Waterloo et al., 2006). The annual average temperature is 26.7 °C and the VPD ranges from about 7 hPa in the wet to 14 hPa in the dry season (Araújo et al., 2002). The annual rainfall (1966–1992) reported in the Ducke Forest Reserve, about 75 km SE from the Campina Reserve, amounts to 2,442 mm, with a standard deviation of 306 mm, with about 27% of the rain falling in the dry season (Waterloo et al., 2006).

3.2.2 Soil CO₂ measurements

Soil respiration was measured in plateau, slope, valley, THF, SHF and bare soil areas (Table 3.1), with for each site a LI-8100 automated soil CO₂ flux system (LI-COR Inc., Lincoln, NE, USA) using a LI-8100-101 long-term chamber (diameter 20 cm). The measurements were taken at random locations, chosen as follows. First, a point on a main straight line of 60 m length was determined by selecting a distance from 0 to 59 m, according to the time in seconds (0–59) found by reading a stopwatch at random. Next, the measurement location was chosen 10 m sideways from this point, with the direction found from a new stopwatch reading, taking left or right as the number was even or odd, respectively.

The LI-8100 system was set up to measure soil CO₂ flux, together with two soil temperature sensors probes (Type E, LI-COR Inc., Lincoln, NE, USA) that were installed at 5 and 10 cm depths and the 24 hours automatic measurements were taking during 2 minutes leaving 30-minute intervals prior to another measurement. Both temperature sensors collected information at the same time frequency as the LI-8100 system. The measurements, were made during four to seven consecutive days at three locations in each plot (about 240 measurements on the same place), then moving to the next area with the ring previously installed inside the measurement area or in a different location inside the Cuieiras or *Campina* reserves. This measurement design provides spatial and temporal variation for each measurement site and for the average values reported for each plot.

The measurements in the *Campinarana* forest from Cuieiras Reserve were performed using another automated system developed by Alterra, The Netherlands. This system consists of two chambers of 25 cm high and 30 cm in diameter made of polyethylene coupled to a CO₂ LI-840 analyzer (LI-COR Inc., Lincoln, NE, USA) placed randomly in a similar manner as above (Table 3.1). This system consists of a chamber with a lid closed by means of an electric motor with pulley and rope. The measurements were taken when the lid closes automatically at programmed measurement intervals. Inside, the chamber had a fan to homogenize the air (Micronel, D341T, Vista, USA). The air entered in the chamber through polyethylene tubing placed at the base and leaves through a similar tube below the chamber lid (upper part). There was also a pressure balancing tube installed in the chamber, which equalizes the pressure between the chamber and the outside pressure. The lid was sealed

Table 3.1: Site and location of measurement, position, soil type, forest, LAI and period of soil CO₂ respiration measurements.

Site	Location	Position	Soil type	Forest type	LAI	Period of Rs measurements
Cuieras	Plateau	2° 36' 31"S 60° 12' 32"W	Ultisols	Evergreen	6.1±0.6 *	August, 3 until November, 6, 2006 and February, 21 until 26, 2008
	Slope	2° 36' 23"S 60° 12' 36"W	Argisols	Evergreen	5.5±0.5	November, 20, 2006 until May, 31, 2007 June, 28 – July, 12, 2007
	Campinarana	2° 36' 17.9"S 60° 12' 75"W	Spodosols	Evergreen	4.9±0.6	August, 28 – September, 17, 2006 November, 20, 2006 – May, 31, 2007
	Valley	2° 36' 11"S 60° 12' 44"W	Spodosols and Gleys	Evergreen	5.8±0.6	February, 8, 2007, November, 13 – December, 13, 2007 July, 24 – August, 05, 2008
THF	2° 35' 25.48"S 60° 1' 52.30"W	Spodosols	Scleromorphic	4.1±0.4	September, 26 – 30, 2007 October, 01 – 11, 2007	
Campina	SHF	2° 35' 27.29"S 60° 1' 49.90"W	Spodosols	Scleromorphic and Vascular epiphytes	2.5±0.6	July, 20 – September, 01, 2007
	Bare soil	2° 35' 26.75"S 60° 1' 50.52"W	Spodosols	Vascular epiphytes and Lichens	–	September, 6 until 26, 2007

* = Data from Marques-Filho et al. (2005)

with two hollow neoprene bands to prevent any external contamination or exchange with outside air. The analyses for each chamber were provided by an Infra-Red Gas Analyser—IRGA, LI-840 (LI-COR Inc., Lincoln, NE, USA) and the soil moisture and soil temperature (5 and 10 cm) were also measured beside of the chambers, and then the values were stored every 30 minutes on the CR10X datalogger (Campbell Scientific, Inc., Logan, USA). Alterra system was well designed as LI-8100 system. We assayed both systems, installing the LI-8100 for a day beside of the Alterra system and the data analyses from both systems accounted similar emissions on that day, with average emission of 5.2 and 4.3 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$, respectively (R^2 of 60%).

Both systems had output in $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$. The measurements were based on closed dynamic chambers (Norman et al., 1992) and the flux was calculated by a linear rise in CO₂ concentration inside of the headspace during each measurement. Following Hutchinson and Livingston (1993), the collars were inserted at 2 cm depth into clayey soils and 3 cm into sandy soils. After inserting the polyvinyl chloride (PVC) rings on the soils, the start of measurements needed to be delayed a certain time to allow a re-stabilization of soil surface, which varied from site to site to detect the outliers of standard measurements as described in Zanchi et al. (2011, 2012). Most of the measurements were done after more than two weeks to one month after ring installation (Bréchet et al., 2011), depending on the soil and forest type (Zanchi et al., 2012). Both systems had been programmed to exclude the first 30 seconds of increase of the CO₂ concentration (“Dead-Band period”), to prevent any wrong CO₂ estimation. Measurements must have a time-concentration fit with an R^2 of 99% at least to be accepted.

The IRGA’s systems were calibrated using the LI-COR calibration protocols. During the field campaign, three calibrations on each IRGA were performed: respectively one, four and eight months after the experiment started. A standard CO₂ gas (512±2% ppm, Praxair, Osasco, São Paulo) was used for the “gain”-calibration, and an ultra-pure nitrogen gas (5.0 analytical, same manufacturer) containing neither CO₂ nor H₂O was used to perform the “zero”-calibration.

3.2.3 Leaf Area Index (LAI) measurements

LAI was measured by a non-destructive method, using a plant canopy analyzer sensor, LI-2000 (LI-COR Inc., Lincoln, NE, USA). For the Cuieiras reserve, valley and *Campinarana* forest the measurements were performed twice a month during April–September of 2007 and March–April of 2008. Another measurement was made on the slope forest during April of 2008. In the *Campina* reserve, the measurements were also made twice a month during May–October of 2007 on the SHF and THF forests (Zanchi et al., 2012). The LAI values were collected using the LI-2000 measurement protocol which must be taken during cloudy days and early morning or later afternoon with 10 measurements below the canopy and one above to the system calculate the

LAI values.

The measurement design consists of two grids of tree parallels lines in the Cuieiras reserve, one located in the *Campinarana* and another in the Valley forest and two others on the *Campina* reserve, one located at SHF and another one at THF. The measurement design used a grid consisting of a nylon rope with three parallel lines of 100 m long and separated by 15 m width between each other. The measurements were made meter by meter and the LAI for each area was averaged over all measurements. The LAI average measurements were obtained through 300 measurement repetitions which decrease the standard error below 0.6 and CV was below 0.4 for all measured sites. For the plateau forest, the values of LAI found by Marques-Filho et al. (2005) were assumed.

3.2.4 Model description

The model (Eq.(3.1)) is run with a time step of 1 hour. We attempt to model the dependence of soil respiration (R_s) on soil temperature (T), with the latter estimated from measurements at two depths. The starting point for model development in this study is the exponential Q_{10} model (van 't Hoff model), but modified and used by Lloyd and Taylor (1994); Fang and Moncrieff (2001); Davidson et al. (1998), as follows:

$$R_s = R_0 * e^{\beta_0 * T} \quad (3.1)$$

where R_0 and β_0 are parameters obtained through the fitting of observed R_s values ($\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) to observed T values ($^{\circ}\text{C}$). This is the traditional method (*i.e.* Howard and Howard (1993); Janssens and Pilegaard (2003); Davidson and Janssens (2006)), which is applied here to diurnal series. Fitting is done using a non-linear least square fitting procedure. The factor Q_{10} by which R_s is multiplied when T increases by 10°C , is determined as:

$$Q_{10} = e^{10 * \beta_0} \quad (3.2)$$

A common problem with this approach is that often a hysteresis is found between the measured diurnal courses of the soil temperature and the respiration efflux. Hence it is interesting to calculate a time lag (t_{lag}). The diurnal cycle of soil temperature $T(z, t)$ can be approximated by a diurnal sine cycle, depending on depth z (cm) and time t (s) according to

$$T(z, t) = T_{mean}(z) + A(z) * \sin(\omega * (t - t_{shift}(z))) \quad (3.3)$$

where $\omega = 2 * \pi / \text{daylength}$ (rad s^{-1}) is the radial frequency, T_{mean} is the average temperature over the period of measurement, $A(z)$ is the temperature amplitude (one-half the difference between maximum and minimum temperatures), and $t_{shift}(z)$ is a time shift (Arya, 1988; Elias et al., 2004; Gao et al., 2008; Sierra et al., 2011). To determine the time lags between temperature

series and the R_s , we also need the phase of the R_s , which is found by fitting a diurnal harmonic to the R_s series itself, yielding the time shift $t_{shift,resp}$. Finally,

$$t_{lag}(z) = t_{shift,resp} - t_{shift}(z) \quad (3.4)$$

This procedure is repeated for each observation depth for each fieldwork site.

We also use the (estimated) surface temperature T_{sur} for modelling, since, often, most of the heterotrophic respiration has its source near the surface, and the autotrophic respiration is determined by above-ground factors. The practical details are as follows. Assuming homogeneous heat conduction, $t_{shift}(z)$ grows linearly with depth, whereas the amplitude $A(z)$ decreases exponentially. Hence, if one has a series of temperature measurements at two depths, one can estimate the course of the temperature at any “reference” depth by fitting first a diurnal harmonic to each of the two, and subsequently interpolating or extrapolating the time shift and amplitude to the reference depth (the amplitude should be extrapolated logarithmically instead of linearly, thus not A but $\log(A)$ which has to be linearly extrapolated). We use this to estimate the temperature series at the surface. $A(0)$, $t_{shift}(0)$ and $T_{mean}(0)$ are determined by extrapolation to $z = 0$. The extrapolation is illustrated by (Fig. 3.2). The resulting T_{sur} is

$$T_{sur}(t) = T_{mean}(0) + A(0) * \sin(\omega * (t - t_{shift}(0))) \quad (3.5)$$

The surface time lag $t_{lag}(sur)$ is subsequently calculated using Eq.(3.4). Then, the parameters of the respiration model are fitted, using now T_{sur} . When fitting, no attempt is made to remove the remaining time lag (which is usually small).

3.3 Results

3.3.1 Soil CO₂ efflux and temperature variations

For all measurement sites, average diurnal time series for soil respiration and temperature (at two levels) were calculated (when possible). This was done for the dry season (July–September) and wet season (other months) separately. Note that the seasons were in general incompletely sampled (see Table 3.1), which will affect the resulting averages to some extent. The following results pertain to these average series. No data are available for some stations in the wet season (Table 3.2).

Whereas the dependence on season is relatively weak, the daily average soil CO₂ efflux varied considerably between the sites, with values ranging from 1 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ in a well-drained site of bare soil at the Campina Reserve (dry season), to 5.6–6.0 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ in the *Campinarana* forest at the Cuieiras Reserve (Fig. 3.3 and Table 3.2). The bare soil site,

Table 3.2: Mean dry and wet period of soil CO₂ efflux (R_s) in $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$, mean of soil temperature ($^{\circ}\text{C}$) at 10 and 5 cm; coefficient of determination (R^2) between R_s and observed (10 cm, 5 cm) or modelled (surface) temperature, time lag (t_{lag}) of respiration with respect to temperature for the given level, in hours, Q_{10} factor for model-fitted respiration using the surface temperature.

Site	Location	R_s	$T_{10 \text{ cm}}$	$T_{5 \text{ cm}}$	R^2 (10cm)	t_{lag} (10cm)	R^2 (5cm)	t_{lag} (5cm)	R^2 (sur)	t_{lag} (sur)	Q_{10} (sur)
Wet Period											
CU	Plateau	2.6±0.2	25.8±0.1	25.8±0.2	0.74	-2.0	0.89	-1.0	0.92	0.0	5.2
	Slope	3.9±0.3	25.7±0.1	25.6±0.2	0.46	-0.2	0.53	0.7	0.56	1.7	3.8
	CPRN	5.6±0.4	25.2±0.1	25.7±0.2	0.00	-6.2	0.36	-3.0	0.81	0.3	3.3
	Valley	4.5±0.5	25.8±0.1	25.9±0.2	0.19	-7.7	0.43	-3.1	0.84	1.4	7.7
CP	THF	3.8±0.2	25.6±0.2	25.7±0.3	0.45	-0.7	0.23	1.3	0.09	3.2	1.3
	SHF	–	–	–	–	–	–	–	–	–	–
	Bare soil	–	–	–	–	–	–	–	–	–	–
Dry Period											
CU	Plateau	2.3±0.1	25.8±0.2	25.8±0.3	0.34	-3.6	0.57	-2.5	0.72	-1.5	2.0
	Slope	3.0±0.1	25.7±0.2	25.7±0.3	0.66	-0.8	0.68	-0.2	0.77	0.5	2.3
	CPRN	6.0±0.6	25.2±0.1	25.4±0.3	0.02	-6.6	0.23	-3.4	0.69	-0.1	2.5
	Valley	3.7±0.5	25.0±0.1	25.0±0.3	0.00	-5.9	0.65	-2.1	0.73	1.8	5.5
CP	THF	3.9±0.2	25.8±0.3	25.9±0.4	0.45	0.5	0.23	2.3	0.09	4.1	–
	SHF	1.4±0.1	25.4±0.5	25.2±0.9	0.39	2.3	0.11	4.5	0.01	6.7	–
	Bare soil	1.0±0.1	28.6±2.2	28.6±2.4	0.95	-0.4	0.96	0.0	0.78	0.4	1.7

CU = Cuieiras; CP = Campina ; THF = Tall heat forest; SHF = Stunted heat forest; CPRN = Campinarana;
 $\pm x.x$ = standard deviation

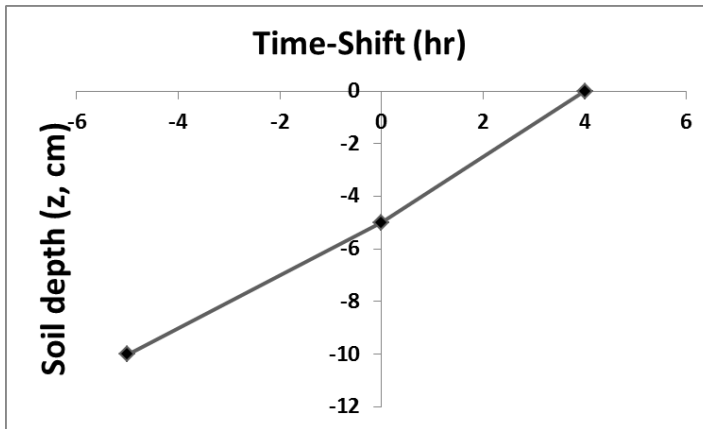


Figure 3.2: Illustration of the depth dependence of the time shift, and its extrapolation to the surface.

which had the lowest emission, had on the other hand a higher average soil temperature at 5 cm (28.6 ± 2.4 °C) than the other sites (25 to 26 °C with small variation), which indicates that there is no straightforward relation between the temperatures and the soil CO₂ emissions across these sites.

The very high average efflux for Campinarana (both seasons) is not due to the fact that a different instrument was used there; a brief instrument comparison at a single site even showed a slightly lower average for this instrument compared to the standard instruments (section 3.2.2).

Soil CO₂ efflux and soil temperature showed a diurnal cycle (Fig. 3.3), but with a site-dependent hysteresis (time lag) between soil temperature and soil CO₂ efflux (Table 3.2). Generally, sites with tall forests showed similar time-series, with increased CO₂ production starting at about 9:00 AM and decreasing at about 7:00 PM, and increased soil temperature at 5 cm around 11:00 AM and decreasing around 11:00 PM (Fig. 3.3a). Deviations of the respiration from a harmonic course occur at various degree, and it is difficult to determine a systematic behaviour in this. For the *Cuieiras* forests, CO₂-emission increased and decreased earlier than for the THF and SHF forests (Fig. 3.3). For the bare soil site, the emission was in phase with the 5 cm temperature (Table 3.2).

Because of the time lags, the temperature data are not immediately useful for the parametrization of the respiration. This is reflected by the low R^2 found when using a simple model without lags (Table 3.2). In the following, we describe how this problem is largely removed by applying the analytical model of section 3.2.4 to the temperature data.

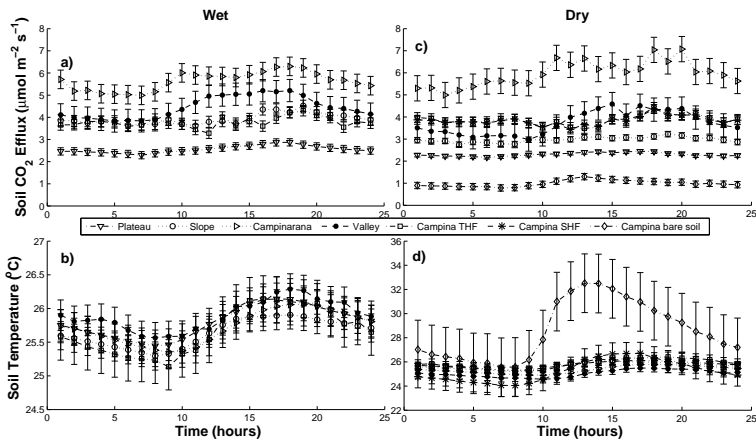


Figure 3.3: Ensemble averages and standard deviations of the diurnal course of the soil respiration (top) and 5 cm temperature (bottom) for the seven Amazonian sites at Cuieiras Reserve (plateau, slope, *Campinarana*, valley forests) and in the *Campina* Reserve (tall heath forest–THF, stunted heath forest–SHF and bare soil areas). Left: a) and b) soil respiration and temperature for wet season, right: c) and d) soil respiration and temperature for dry season.

3.3.2 Modeling of soil respiration to the temperature response

Eq. 3.1 was fitted to the data, using the observed T from 5 cm and from 10 cm. The observed and modelled respiration are compared in Fig. 3.4 for the Cuieiras sites, and in Fig. 3.5 for the Campina sites. The corresponding coefficients of determination R^2 are shown in Table 3.2.

Many plots in Fig. 3.4 show a similar behaviour. Since the temperature at 5 and 10 cm lags behind the observed respiration, this is also the case with the modelled respiration. Moreover, the phase difference causes a diminishing of the modelled amplitude. If the phase lag is about a quarter period (6 hours) or more, which is often the case for 10 cm temperatures, the amplitude may go to practically zero, or even a negative correlation may be found between the temperature and the modelled respiration. To illustrate this phenomenon, Fig. 3.6 shows observed and modelled respiration against temperature. Figure 3.6a shows a typical case with modest hysteresis (time lag about -2.0 hours), but Fig. 3.6b is a case where the 10 cm hysteresis is so strong (time lag about -7.7 hours) that the slope of the best fit has become negative (which often occurs). Thus, it is obvious that when diurnal cycles are used for fitting, optimized model parameters such as Q_{10} are often not meaningful if no correction for the hysteresis is made.

In an attempt to remedy the problems caused by the time lag, extrapolated surface temperature series were constructed using Eq. 3.5. Fig. 3.7 illustrates

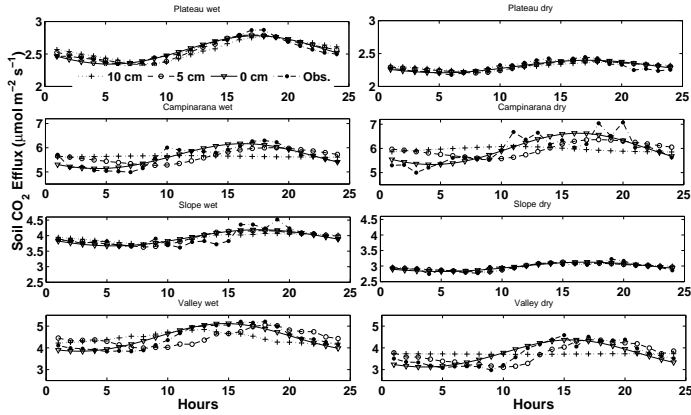


Figure 3.4: Ensemble-averaged diurnal course of the soil CO₂ efflux for the four Cuieiras sites. Modelled with temperature for 10 cm and 5 cm depth (observed) and of 0 cm depth (extrapolated), and observed values. Left: wet season, right: dry season.

how the time lags are extrapolated. Table 3.2 shows the resulting time lag ($t_{lag}(sur)$) for the surface. For the Cuieiras and bare soil sites, the surface time lags have a relative narrow distribution around zero, with all deviations below 2 hours, in contrast with the time lags at 5 and 10 cm which are in general negative and much more variable. The small time lags show that for these sites the respiration is often closely in phase with the surface temperature unlike the measured temperatures. For SHF and THF on the other hand, large positive time lags are found when using the surface temperature, but the figures indicate that the observed respiration series are so irregular that it is difficult to define precise time lags. For those sites it is hard to model the respiration from the surface temperature alone.

Figure 3.8 shows the average time lag against the LAI-values for the sites. Low LAI-values (related to nutrient-poor soil) correspond to long time lags, whereas for high LAI values (nutrient-rich) the time lag tends to disappear. However, a robust quantitative relation is hard to infer, as especially the sites with low LAI are very low in number, and their time lags cannot be sharply determined, owing to the relatively large noise in the respiration data.

The respiration model was fitted to the surface temperature, and the results have been added to Fig. 3.4 and Fig. 3.5, and Table 3.2. For most measurement locations, the R^2 between average soil CO₂ efflux and soil temperature was better for the 5 cm than for the 10 cm temperature (Table 3.2), and R^2 was usually further improved with the surface temperature. Figure 3.6 illustrates the diminishing of the hysteresis, which shows a great performance for cases with strong hysteresis at the observation depths (Fig. 3.6b). For the Cuieiras

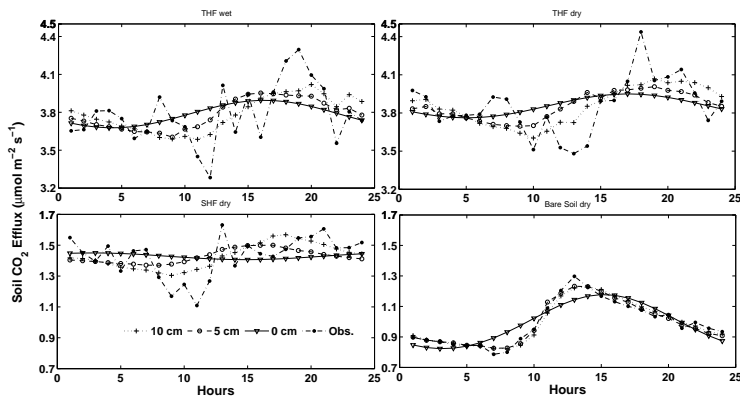


Figure 3.5: Ensemble-averaged diurnal course of the soil CO_2 efflux for the Campina sites: THF site (wet and dry season), SHF and bare soil sites (dry season, no other measurements available). Modelled with temperature for 10 cm and 5 cm depth (observed) and of 0 cm depth (extrapolated), and observed values. Top left: wet season, others: dry season.

sites, the coefficient of determination R^2 increased from on average 0.54 for 5 cm to 0.72 for the T_{sur} . However, R^2 did not improve for SHF and THF, because of the irregular respiration data series. For the bare soil site, the R^2 at 5 cm was the highest of all (0.96), so that application of the sinusoidal model did not improve the correspondence. On the other hand, the model adjusts the temperature series into a sinusoidal shape, whereas both the real temperature and the respiration series have very non-sinusoidal shapes for this site. This is typical for a soil exposed to direct insolation. As a consequence, the R^2 between the modelled and observed respiration decreased to 0.78.

The Q_{10} -factor depends on the depth of the T measurements. An increase with depth is to be expected because the amplitude of the temperature cycle decreases, though for the present diurnal modeling it may turn into a decrease because of the growing phase difference (Fig. 3.4 and Fig. 3.5). The good phase-correspondence between respiration and surface temperature for many sites, and the conjecture that much of the heterotrophic respiration takes place near the surface, are reasons for choosing the Q_{10} which was found using the surface temperature. The values for the Cuieiras sites and the bare soil site are shown in Table 3.2. For SHF and THF, no values are shown since the very low R^2 (at most 0.11, against 0.56 at least for the other sites) makes it difficult to estimate a Q_{10} value there.

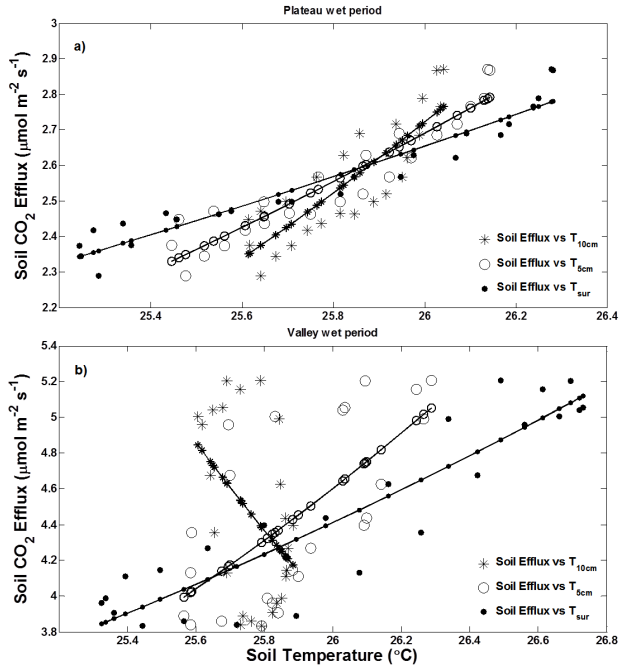


Figure 3.6: Soil CO₂ efflux versus soil temperature in the Cuieiras reserve. The stars (*), open dots (o) and closed dots (●) represent ensemble-averaged hourly values pertaining to soil temperatures at 10 cm, 5 cm, and the surface, respectively. The loose symbols are observed, the symbols on the curves are modeled averages. a) plateau wet period and b) valley wet period.

3.4 Discussion

The observed diurnal and seasonal variations in the temperature and respiration were in general quite small, which is typical for the tropical regions to which the present results are restricted. Hence, it is uncertain whether the observed soil temperature–respiration relations (e.g. the Q_{10} values) can be extrapolated to e.g. mid–latitude regions where climates with a much larger temperature range prevail. This should be kept in mind when using the results.

Further, when interpreting the results it is important to consider uncertainties in the measurements. The soil temperature measurements are more smooth, when compared with the respiration measurements. The latter are noisy, and as is obvious from the diurnal time series (Figs. 3.4 and 3.5), the noise amplitude is strongly dependent on the site, and highest for the nutrient–poor forest sites (SHF and THF)

A consistent positive relationship between soil respiration and different measures of plant production has been found in different studies (Raich and Schlesinger, 1992; Metcalfe et al., 2011). This is expected as the output of carbon should be in the long term in balance with the input (unless lateral transport and long–term changes in the carbon pool are important). The low productivity of the *Campina* sites explains the low respiration for SHF and bare soil. All *Campina* sites have a lower LAI than the Cuieiras sites (Table 3.1). These sites have small patches of vegetation or no vegetation at all (Luizão, 1996), and the plants are growing slowly compared to the Cuieiras reserve. However, though the base respiration is low, the daily maximum respiration can be relatively high, because of the high temperatures through the daily insolation exposure.

On the other hand, Cuieiras Reserve showed high LAI–values (Table 3.1), indicating adaptation to a rapid resource acquisition with high GPP (Metcalfe et al., 2011), which corresponds to a high average soil respiration (Fig. 3.9). However, though the THF site had a lower LAI than all the Cuieiras sites, and species with slower growth, it had a respiration which is comparable to the Cuieiras average. Fig. 3.9 indicates that there is a tendency of high soil respiration for high LAI, but there is no clear relation between LAI and average soil respiration (R^2 of 41%).

Soil respiration showed a large variation between the Cuieiras sites, increasing from highest (plateau) to lowest (Campinarana, Valley) elevation (Table 3.2). This likely reflects differences in the topography of the ecosystem, nutrients and C sources, and species adaptation (Metcalfe et al., 2011). First, there is a difference in soil type: clayey for the plateau and slope (Luizão et al., 2004) but sandy for valley and *Campinarana*. The *Campinarana* forest is located on foot of a slope, and has more soil organic carbon content, because of the fine root mat and fluctuation of phreatic level that provides and keeps more carbon in the upper layers than for other forests (Nepstad et al., 1994; Luizão, 1996; Marques, 2009). For the valley forest, the soil is also sandy, and most of time the groundwater level is near the surface (Zanchi et al., 2011). For this

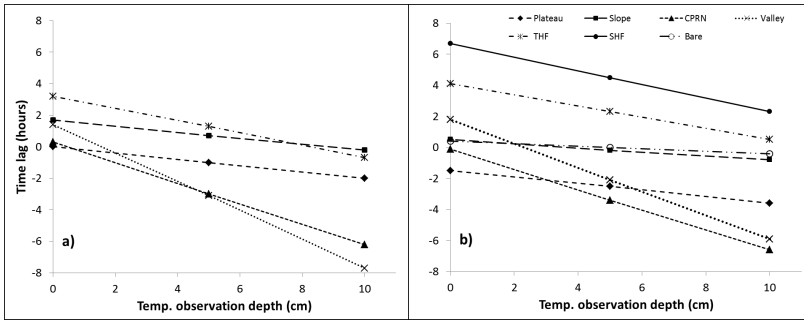


Figure 3.7: Time lag in hours at the surface in dependency of the soil temperature observation depth (cm). a) wet period and b) dry period.

site, CO₂ release is apparently suppressed in comparison to *Campinarana*, because a high phreatic level provokes a slow diffusion of oxygen, often allowing only anaerobic decomposition, with generally slower degradative enzymatic pathways (Linn and Doran, 1984; Davidson and Janssens, 2006). However, the soil CO₂ emission for the valley forest is still high.

For the time lags between temperature and respiration, a great variation is also found between the sites (Table 3.2), but we saw that the variation is strongly reduced when extrapolated surface temperatures are used. Subke and Bahn (2010) also tested the hysteresis between $T-R_s$ for many soil depth-measurements, finding a better fit close to the surface. Our data indicate that soil respiration for the Cuieiras forests and the bare soil site is closely in phase with the soil temperature at the surface, whereas for the *Campina* forests SHF and THF, considerable time lags (3–7 hours) are found. This seems to indicate that for this region, long time lags are typical for scleromorphic vegetation growing on nutrient-poor soil, and do not occur for nutrient-rich soil, neither for bare soil. The number of sites with substantial time lags is so small (two, THF and SHF) that it is difficult to say by what factors they are determined. Based on the literature, one could speculate that the time lag for the scleromorphic vegetation is related to the slow translocation mechanism of photosynthate to the roots and associated microbes which has been reported for species with slow growth rate (Baldocchi et al., 2006; Metcalfe et al., 2011). Unfortunately, we have no data to test this hypothesis. Figure 3.8 shows that the time lag is larger for sites with lower LAI, reflecting scleromorphy, but there are insufficient data and sites to investigate this further. However, uptake of root exudate by microbial processes can be excluded as a cause as it is a fast process (Blagodatskaya et al., 2009), and heterotrophic respiration follows the local temperature cycle rapidly (Tang et al., 2005; Subke and Bahn, 2010).

Our data (Table 3.2) show that the Q_{10} values (with respect to the surface temperature) had a large variation from site to site: 1.3 to 7.7, with the smallest value (< 2) for the bare soil site, intermediate values (dry season: 2.1 to

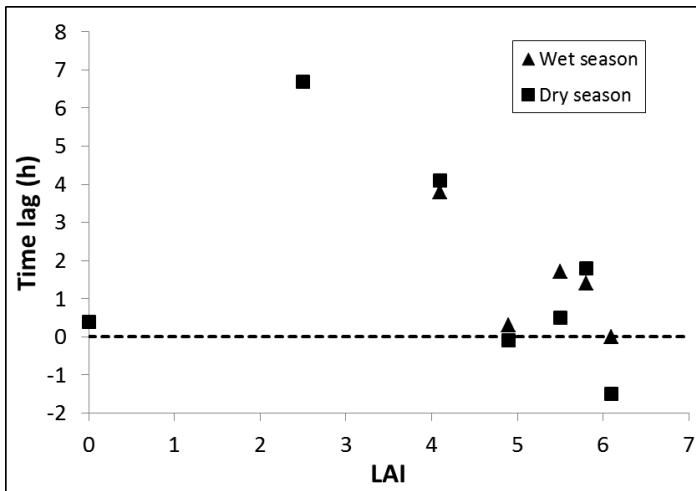


Figure 3.8: Average of LAI versus time lag (with respect to surface temperature) in hours for each forest for the wet and dry season.

2.5, wet season: 3.3 to 5.2) for the Cuieiras sites except valley, and the highest values (dry and wet season: 5.5 and 7.7) for the valley site (for SHF and THF, Q_{10} was difficult to estimate because R^2 is very small). These three groups can be characterized as nutrient-poor (with open vegetation), nutrient-rich, and swampy, respectively. Note also that generally the values of the Q_{10} were higher in the wet than in the dry season. It should be considered that the Q_{10} values and even their ranking might shift somewhat if we would have averages for the complete seasons, which are not available now. But all the results suggest that Q_{10} increases with soil wetness. This may point to the heterotrophic component of soil respiration. Compared to this, root respiration and root-associated respiration were reported to be relatively insensitive to temperature (Heinemeyer et al., 2007), as these components are more dependent on photosynthesis and on the transport of its products (Baldocchi et al., 2006). However, since these latter factors correlate with soil temperature, the actual reason for the strong dependence of Q_{10} on the ecosystem remains unclear (Baldocchi et al., 2006; Kuzyakov and Gavrichkova, 2010; Phillips et al., 2011; Savage et al., 2013).

It is noteworthy that for the bare soil, the uncorrected T at 5 cm correlated very well with the respiration. Applying the corrections leads to a worse correlation (only for this site). It appears that for this site the respiration closely followed the soil temperature, but since both series have a very disharmonic diurnal course (because of a lack of shading), the corrections disturb the good relation by forcing the temperature into a harmonic shape which is unnatural for this site.

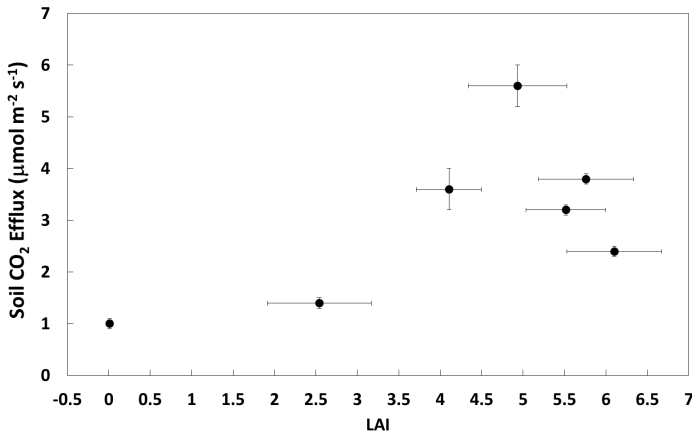


Figure 3.9: Average of LAI versus soil respiration for each forest. The bars are the standard errors.

3.5 Conclusions

Use of the extrapolated surface temperature led to a smaller time lag between temperature and respiration for the Cuieiras Reserve sites than the use of temperatures measured deep in the soil. Consequently, the R^2 between observed and modeled respiration also improved. Interestingly, there was also a clear grouping of the (site-dependent) respiration parameters when surface temperatures were used.

The strongest respiration occurred for the wettest and lowest lying nutrient-rich sites. The tall heath forest, although nutrient-poor and with slowly growing vegetation, had a higher respiration than some nutrient-rich sites. Time lags between surface temperature and respiration were small for the nutrient-rich, but large for the nutrient-poor forests. For the bare soil sites, there was no phase difference. Sensitivity of the respiration to temperature was highly variable, lowest for the nutrient-poor sites, and highest for the wet nutrient-rich sites. Temperature sensitivity was also higher in the wet than in the dry season.

The present work shows that for the considered region, respiration modeling based on (extrapolated or measured) surface temperature can often be recommended. It also indicates the typical dependence (which is found even on a small scale) of modelling parameters such as Q_{10} on ecosystems and season. However, much about the site-dependence remains unexplained. We also note that the analysis was restricted to a daily time scale, and consequently the results cannot be straightforwardly extrapolated to modelling along the year. The present results may offer some guidelines for the design and interpretation of further measurements on soil respiration in the Amazon forest.

Chapter 4

Water balance, nutrient and carbon export from a heath forest catchment in central Amazonia, Brazil³

4.1 Introduction

The Amazon forest extends into eight Latin American countries and covers an area of approximately $6 \cdot 10^6$ km². It is among the largest and most important ecosystem for carbon storage in the world and plays an important role in both regional and global water and carbon cycles (Marengo, 2006; Neill et al., 2006; Cox et al., 2004). Its vastness, spatial variation in forest and soil types and recent changes in land use complicate the quantification of water, carbon and nutrient fluxes in the basin (Marengo, 2006; Raymond, 2005; Fearnside, 1993). In many instances no reliable data are available for large regions covered by different forest types (McClain and Richey, 1996; Remington et al., 2007; Marengo, 2005). In recognition of the importance of water, carbon and nutrient dynamics for the Amazon rainforest, studies have been carried out to estimate fluxes at micro-catchment to basin scales (Leopoldo et al., 1982; Franken and Leopoldo, 1984; Marengo, 2006; McClain et al., 1997; Houghton et al., 1998, 2000; Andreae et al., 2002; Richey et al., 2002; Moreira-Turcq et al., 2003a,b; Telles et al., 2003; Waterloo et al., 2006; Biggs et al., 2006; Neill et al., 2006; Tomasella et al., 2007; Remington et al., 2007). Estimates

³The contents of this chapter have been submitted as F. B. Zanchi, A. P. Tapia, M. J. Waterloo, M. S. Alvarado Barrientos, M. A. Bolson, F. J. Luizão, A. O. Manzi and A. J. Dolman. Water balance, nutrient and carbon export from a heath forest catchment in central Amazonia, Brazil, submitted to Hydrological Processes

for the water balance and hydrological nutrient and carbon are available and spatial and temporal variations have been recognised (Marengo, 2006), these are almost exclusively based on integrative micro-meteorological and catchment water balance studies and do not provide information for specific vegetation types in the terra firme rainforest ecosystem, such as riparian forest or heath forest. Nutrient fluxes are generally low in central Amazonia (Lesack, 1993; Furch and Junk, 1997), but the export of carbon from the rainforest ecosystem by rivers can be significant and often originates from a zone along the stream (Waterloo et al., 2006). Hydrological export of carbon is high in valley or heath forest areas in central Amazonia that are source areas of black water streams – streams high in DOC (McClain and Richey, 1996; Waterloo et al., 2006).

In Central Amazonia, the landscape consists of well-drained, rather flat plateaus separated by steep slopes and broad poorly-drained river valleys (Chauvel et al., 1987; Waterloo et al., 2006). Several forest types have developed in this landscape in response to geology- and topography-controlled distribution of soil types (clayey *versus* sandy soils), inundation and soil drainage status and nutrient availability (Singer and Aguiar, 1986; Walker, 1987; Laurance et al., 1999; Waterloo et al., 2006; Nobre et al., 2011). In general, tall *terra-firme* rainforest covers the well-drained clayey plateaus and slopes, whereas a somewhat shorter-statured forest (locally called campinarana - ecotone forest) grows in the poorly-drained sandy valleys. Using remote sensing images for forest classification in a 18,000 km² terra firme forest area in central Amazonia, Nobre et al. (2011) demonstrated that poorly-drained sandy valley areas covered by campinarana or ecotone forests occupied up to 58% of the area. Low-statured heath forests (*campinarana* and *campina* forests) grow under high water and nutrient stress conditions on upland areas with strongly leached white sandy quartz soils and relatively low phreatic groundwater levels. These cover about 6% of Amazonia (McClain et al., 1997; Luizão et al., 2007). Topography has also been demonstrated to cause spatial variation in nighttime and early-morning atmospheric CO₂ concentrations that may influence photosynthesis and transpiration rates by forests in the valleys of central Amazonia (Araújo et al., 2008a,b; Araújo, 2009). The presence of such a variation of functional forest types within short distances in the Amazonian rainforest ecosystem may affect the overall sensitivity of its water, nutrient and carbon cycles to climate change that cannot be studied through larger-scale integrative micro-meteorological or catchment water balance and nutrient flux studies.

Precipitation over the Amazon is predicted to decrease (Betts et al., 2008) and a recent modelling study also suggested a significant decrease in runoff for Brazil (Arnell and Gosling, 2013) due to future climate change. A decrease in precipitation would reduce recharge on the plateaus thereby causing a permanent lowering of the groundwater levels in the poorly-drained valleys that are now covered by campinarana – ecotone forest. In a drainage experiment, Zanchi et al. (2011) demonstrated that a permanent lowering of the groundwater level in the valley of the Asu catchment led to sharp decreases

in soil respiration and soil C and N content, which approached the low levels observed in upland campina forest. It is conceivable that a permanent lowering of the phreatic level in these valleys could in the long-term trigger a change from poorly-drained campinarana – ecotone forest to better drained campinarana – campina forest. Such a change could have implications for the supply of moisture to the atmosphere when evapotranspiration rates decrease, and may impact carbon and nutrient fluxes into the river system. At present, little is known about evapotranspiration, carbon and nutrient dynamics of upland *Campina* and *Campinarana* forests (Becker, 1996b; McClain et al., 1997; Tyree et al., 1998; Vernimmen et al., 2007). Furthermore, many of the parameters needed to configure land surface platforms of atmospheric models to study feedbacks on the water cycle of land cover change are not very well known (Avissar et al., 2002). Understanding the water balance, together with transpiration and evaporation dynamics of Amazonian heath forests, is therefore an important step towards more realistic and fine scale modelling of the carbon dynamics of the Amazonian basin under future scenarios of climate and land cover change (Avissar et al., 2002; Gash et al., 2004).

The objective of the present study is to quantify the components of the annual water balance and exports of dissolved carbon and inorganic macronutrients from upland campina/campinarana forest to contribute to a better understanding of the implications that a change in valley water level and vegetation cover, as described above, might have. To achieve this, the annual water balance of a micro-scale campina and campinarana heath forest catchment in central Amazonia was quantified by combining catchment water balance measurements with sapflow transpiration measurements, micro-meteorological evapotranspiration and rainfall interception measurements. Hydrological carbon and macro-nutrient (N, P, K, Ca, Mg) exports were subsequently quantified using regular hydrochemical analysis of surface and groundwater. The latter had not been quantified before for campina and campinarana forest in the Amazon, or elsewhere.

The study was done within the framework of the Large Scale Biosphere–Atmosphere Experiment in Amazonia (LBA), an international research initiative led by Brazil that started in 1997 (Nobre et al., 1997; Avissar et al., 2002).

4.2 Site description

The *Reserva Biológica de Campina*, managed by the *Instituto Nacional de Pesquisas da Amazônia*–INPA, covers approximately 900 ha of largely undisturbed lowland evergreen rainforest. This reserve includes 6.5 ha of stunted campina heath forest and taller campinarana heath forests. The Campina catchment (2°35'30.26 S, 60°01'48.79 W, 93–101 m a.s.l.) is located in the reserve at about 60 km north of Manaus, AM, Brazil (entrance at km 43 along the BR–174 highway). Both campina and campinarana forests grow on well-drained, strongly leached white sandy soils on rather flat plateaus in the

landscape. Due to the flatness of the area it proved difficult to estimate the catchment area from 30 m Shuttle Radar Topographic Mission (SRTM) digital elevation data. Three tracks were therefore made to locate and verify the Shuttle Radar Topographic Mission derived watershed position visually using a Garmin GPSMAP 76CSx leading to a catchment area estimate of 7.5 ha. The campina area covered 2.6 ha and consisted of a mosaic of trees, with a canopy height of 4–7 m, and non-vegetated areas where the white sandy quartz soil was exposed. The leaf area index (LAI) was 2.4 ± 1.3 . It was surrounded by 3.9 ha of campinarana forest with a canopy height of 10–18 m and an LAI of 3.9 ± 1.0 (Zanchi et al., 2012). The lowland terra firme rainforest in the remaining area of the reserve has a canopy height of 25–35 m and grows on more clayey soil (Luizão, 1996; McClain et al., 1997).

Heath forests have low species richness, dominated by one or more species and unusual physiognomy: shorter stature, many-branched and tortuous trees and bushes with scleromorphic leaves and a considerable load of vascular epiphytes (*Orchidaceae*, *Bromeliaceae*, *Araceae*, *Ericaceae*) and lichens (Anderson, 1981; Richards, 1996). The main shrub and tree species in central Amazonian heath forest are *Ouratea spruceana* (*Ochnaceae*), *Pagamea duckei* (*Rubiaceae*), *Pradosia schomburgkiana* (*Sapotaceae*) and *Aldina heterophylla* (*Caesalpiniaceae*) (Anderson, 1981; Luizão, 1996). In addition to the common species mentioned above, there are also endemic species which belong only to these areas due to adaptation to the specific and harsh soil, microclimate and environmental characteristics.

The Campina catchment included Oxisols and Spodosols (McClain et al., 1997). The former are in the riparian zone, whereas the latter are on the plateau where the vegetation is low-statured Campina forest and on the hill slopes covered by taller campinarana forest. The coarse grained Spodosol is poor in nutrients, because of rapid excessive drainage and its low capacity to retain nutrients (Prance and Schubart, 1978; Luizão, 1996). The stream was fed by a source in a small valley incised in the sandy plateau. The soils on the lower slopes seem to be of somewhat finer texture than the sandy soil on the plateau.

The Asu catchment is a primary rainforest catchment in INPA's *Reserva Biológica de Cuieiras* about 19 km to the west of the Campina catchment. The main forest types in this catchment are tall terra firme rainforest on the well-drained clayey plateaus and slopes, campinarana – ecotone forest on the sandy foot slopes in the valleys and riparian forest on a broad water-logged area along the stream channel. More details about this catchment were published in Araújo et al. (2002), Waterloo et al. (2006), Tomasella et al. (2007) and Araújo et al. (2010).

The climate in the region does not show much seasonal variation. The annual average temperature is 26.7°C (McClain et al., 1997) and relative humidity is high at about 80%. Long-term (1927–2007) annual average rainfall in Manaus, about 44 km SE from the Campina reserve amounted to 2211 ± 310 mm (Satyamurty et al., 2010). The dry season is from June to November but monthly rainfall may still exceed 100 mm in this period. During dry periods,

the *Campina* forest may experience more frequent drought stress than the rain-forest in the Asu catchment because of the low plant available water capacity of the coarse sandy soil and the fast drainage to groundwater after rainfall events.

4.3 Methods

4.3.1 Field measurements

Evapotranspiration

Micrometeorological measurements were made with instruments mounted on a 9.5 m tower in the lowest part of stunted heath forest spanning 366 days from March 18, 2007 until March 18, 2008. The tower extended 3–4 m above the canopy (average height 6 m). Incoming and reflected solar radiation and net radiation were measured with two pyranometers (Skye SKS 1110, UK) and a net radiometer (REBS Q7, USA), respectively. Temperature and relative humidity were measured with Vaisala HUMICAP sensor (HMP35, Finland). Precipitation was measured at canopy level with a tipping bucket rain gauge (EM ARG–100, UK). A second tipping bucket rain gauge (VU University, The Netherlands) was used as a back up for periods when the meteorological set-up was not functioning. Missing data for rainfall were filled in with those measured in the Asu catchment (Waterloo et al., 2006), while realizing that there may have been differences due to the distance between study sites (Asu catchment is approximately 19 km to the West from *Campina* catchment). Soil heat flux was obtained from soil thermocouple temperature measurements made at 0.01 and 0.03 m below the surface. A leaf wetness sensor (SWS-02, Delta-T Devices, UK) was installed at canopy level. The phreatic level was measured close to the micro-meteorological tower with a pressure transducer (Druck PDCR–830, Germany) in a piezometer with its screen at 2.5–3.0 m below the soil surface.

A thin-wire (0.05 mm) Co/Cr thermocouple was installed at the top of the tower for sensible heat flux and actual evapotranspiration measurement under dry canopy conditions (E_d) according to the Temperature Variance Energy Balance (TVEB) method (Tillman, 1972; de Bruin et al., 1993; Vugts et al., 1993; Giambelluca et al., 2009).

A soil moisture content profile was measured with four soil water content Frequency Domain Reflectometers (FDR, CS-615, Campbell Scientific, USA) installed next to the meteorological tower. Sensors were installed horizontally at depths of 0.05, 0.10, 0.30 and 1.00 m below the soil surface.

Sapflow velocities were measured according to the Granier method with permanently heated needles (Granier, 1987, 1985). The 24 sapflow sensors were placed at 2 cm depth in the sapwood of 12 tree stems, with diameters varying between 6 and 36 cm, covering variations of sapflow velocities within

trees, as well as between trees. Sapflow velocities were converted to transpiration rates based on the measured velocities, sapwood and basal areas of the measured trees and the basal area of the Campina forest (Alvarado-Barrientos et al., 2013).

All instruments were connected to Campbell Scientific 21X data loggers (USA) and sampled over 30 s intervals, with the exception of the fast-response thermocouple that was sampled at 0.5 s intervals. Averages and rainfall total were calculated for 10-minute intervals. Averages and standard deviations of the fast-thermocouple temperature were calculated at 5-minute intervals to minimize the influence of trends in temperature on the standard deviation (de Bruin et al., 1993; Vugts et al., 1993).

Actual evaporation under wet canopy conditions (*i.e.* rainfall interception loss E_i) was estimated as the difference between above-canopy rainfall and throughfall (TF) measurements and stemflow (SF) estimates ($E_i = P_G - TF - SF$). Throughfall was measured in two campina and campinarana plots in the *Reserva Biológica de Campina* between May 10 and July 4 in 2007 (Alvarado, 2007). Both plots had been the subject of previous ecological studies (Ferreira, 1997; Luizão et al., 2007). In each plot three parallel 60 m transects were made with numbered positions at 1 m intervals. The roving gauge method (Lloyd and Marques-Filho, 1988) was implemented, such that 30 of 180 possible positions at each plot were selected randomly with the help of a previously generated list of random numbers. Therefore, 10 gauges per transect were positioned randomly and relocated after each sampling occasion. Sampling intervals ranged from one to three days. The gauges were custom made out of 2 L plastic bottles by cutting out and turning upside down the upper part to make a funnel that had a diameter of 0.10 m. The funnel was leveled at approximately 0.15 m above the soil surface. Stemflow (SF) was assumed at a 1.3% of gross rainfall P_G , based on values presented in literature for heath and rainforests elsewhere (Cornu et al., 1998; Cuartas et al., 2007; Vernimmen et al., 2007; Herrera, 1979). The analytical rainfall interception model (Gash, 1979) was used to model interception loss from daily rainfall inputs. The canopy capacity for the campina forest was estimated at 1.15 mm, the free-throughfall coefficient at 0.18 and \bar{E}/\bar{R} at 0.11 (Alvarado, 2007). Rainfall directed to the trunk was taken as 0.015 and the trunk storage capacity at 0.02 mm.

The reference evaporation E_0 was calculated for comparison with measured evapotranspiration rates according to the FAO method using the Penman-Monteith equation (Allen et al., 1998).

River and groundwater data

The Campina catchment discharge (Q) was determined from 5-minute averages of the river water level (H), measured between 18 March 2007 and 19 March 2008. The water level was measured with a compensated pressure transducer (Druck, PDCR-830, Germany) with a 175 hPa range, referenced

to water level readings from a staff gauge. Electrical conductivity (EC) and temperature (T) were measured at the same location with a flow-through electrical conductivity and temperature probe (CS547A-L, Campbell Scientific, Inc., USA). These data were checked against regular spot measurements of EC and T from a calibrated EC-meter (WTW cond340i, Germany). Measurements of Q at different stage heights (0.13–0.24 m) were made on 34 occasions using the salt dilution method (Aastad and Sogren, 1954) to construct a stage-discharge relationship ($R^2=0.82$).

Discharge data were missing for 73 days when the data logger was inoperative. The *HBV light* (Seibert, 2005) rainfall runoff model, calibrated against the measured rainfall–runoff time series, was used to fill gaps in discharge data. Missing EC values were substituted with values obtained from a polynomial fit between discharge and corresponding EC measurements ($R^2=0.83$).

Monthly groundwater discharge (Q_g) was determined from the catchment water balance ($Q_g = P_g - ET - Q_s \pm \Delta S$), where actual evapotranspiration (ET) was the sum of dry canopy evapotranspiration E_d , as obtained from the TVEB method (Tillman, 1972; Vugts et al., 1993), and the rainfall interception loss E_i during wet canopy conditions. Monthly changes in soil moisture and groundwater storages (ΔS) were obtained from soil moisture and phreatic level measurements discussed below. Runoff and groundwater samples were collected at monthly intervals for laboratory analysis.

Soil moisture and groundwater sampling

Seven piezometers were installed in February, 2007, along a 182 m North–South transect from the campinarana forest near the stream to the campina forest near the micro-meteorological tower. Six piezometers (PC1–PC7) were installed at distances of 1, 6, 10, 20, 30 and 36 m from the stream, representing a valley – slope – plateau campinarana forest section, whereas the 7th piezometer was in the middle of the plateau in true campina forest at the meteorological set-up. These piezometers were installed in line with those of McClain et al. (1997). The piezometers had their 0.5 m screens installed down to a maximum depth of 3.8 m depth below the surface on the plateau. During the measurement period the screens always remained below the minimum dry season phreatic level.

To allow comparison of the groundwater chemistry, piezometers were also installed in the Asu rainforest catchment (Zanchi et al., 2011). One piezometer was installed close to the stream in a control plot in the valley forest in the riparian zone where the groundwater remained close to the surface at all times, whereas a second one was installed in the campinarana – ecotone forest where the phreatic level was at about 0.8 m below the surface.

Groundwater sampling for chemical analysis was done from piezometers after flushing their volume three times to obtain fresh groundwater (Domenico and Schwartz, 1990). Sampling occurred at monthly intervals.

Three sets of six vacuum-type ceramic cup soil water samplers (suction

lysimeters) were installed at depths of 0.1, 0.2, 0.3, 0.4, 0.5 and 0.6 m in the *campina* forest plots to obtain samples of soil moisture and shallow groundwater. Before installation, all samplers had been rinsed with a diluted HNO_3 solution ($\text{pH}=1$) to remove absorbed ions and had been flushed with distilled water until the EC of water used for flushing fell below $2\mu\text{S cm}^{-1}$. Samples from the three sets of cup samplers in each plot were bulked, yielding single samples for each depth after each sampling event. One set of samplers was installed below a tree in soil covered by a litter layer, where litter decomposition and root activity could release DOC in soil water. The second set of samplers was placed in bare sandy soil on the plateau and the third set was in between the first and the second sets of samplers close to the vegetated patch with sparse litter cover. A suction of 600 hPa ($\text{pF}=2.8$) was applied to each sampler and the collected moisture was sampled after three days. Bulk samples were collected for each depth, using water available in any of the samplers. The fact that all of the samplers did not always contain water at sampling, caused spread in the measured concentrations due to in- or exclusion of water from the bare soil samplers (low concentration), or those placed under the tree (high concentration).

4.3.2 Laboratory analysis

Water samples collected in the piezometers and suction lysimeters were analysed for Dissolved Organic Carbon (DOC), Dissolved Inorganic Carbon (DIC) and inorganic ions. DOC and DIC samples were filtered *in situ* with pre-ashed glass fibre filter ($0.7\mu\text{m}$ Whatman GFF, USA) and then stored in a 25 ml glass bottle with a Teflon lid to prevent contamination, the bottles were pre-washed with dilute (5%) HCl then thoroughly rinsed with nanopure de-ionized water. HgCl_2 (at $300\mu\text{M}$) was added for preservation. The samples were stored on ice in coolers immediately after collection and were transported to the laboratory at INPA on the same day. DOC and DIC concentrations were analysed shortly after arrival on a TOC analyser (TOC5000A; Shimadzu, Japan) in the laboratory at INPA.

Separate samples were collected for inorganic ion concentration analysis. Samples were filtered *in situ* using cellulose-acetate filters ($0.47\mu\text{m}$, Whatman GFF, USA) and stored in 100 ml high-density polyethylene bottles with a few drops of a 100 mg l^{-1} Thymol solution added for preservation. Samples were transported to the laboratory at INPA and stored refrigerated. The analyses of major cations (Na^+ , K^+ , Mg^{2+} , Ca^{2+} , NH_4^+) and anions (Cl^- , SO_4^{2-} , PO_4^{3-} , NO_3^- , NO_2^-) was done within a few days after sampling on a Dionex ion chromatograph (DX500, USA).

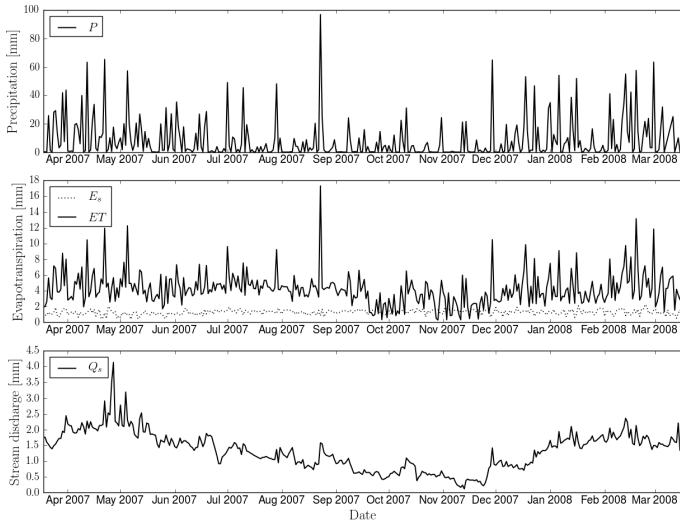


Figure 4.1: Time series of daily rainfall (P), stream runoff (Q_s), transpiration (E_s) and actual evapotranspiration (ET) from the *Campina* catchment.

4.4 Results

4.4.1 Catchment water balance

Rainfall

An overview of the water balance components for the study period is given in Table 4.1. Precipitation amounted to 3054 mm, or 8.3 mm d^{-1} , in the period 18 March 2007 until 18 March 2008 (366 days), with 35 days of missing data substituted by precipitation measured in the Asu catchment in the Cuieiras Reserve. The rainfall total at this site agreed very well with that measured at Asu (2669 *versus* 2643 mm in 331 days). There were 80 dry days, 181 rainy days with daily rainfall less than 10 mm, 78 days with rainfall between 10 mm and 50 mm, and 10 days with rainfall exceeding 50 mm. The maximum daily rainfall was observed in August 2007 and amounted to 97 mm (Fig. 4.1).

The wet season in 2007 was most intense from March to May. In this period only five dry days occurred and none in April (Table 4.1). This was followed by a rather wet period during the dry season in July and August. The driest periods in 2007 were therefore in June and from September to November. Monthly rainfall only once dropped below the corresponding reference evapotranspiration in September 2007. The first two months in the wet season of 2008 were again very wet. Total rainfall in the dry season (June–November)

Table 4.1: Overview of the water balance components for the Campina forest catchment over the period March 18, 2007, until March 18, 2008. Presented are monthly and annual totals of rainfall (P), number of dry days (n), transpiration (E_s), dry canopy evapotranspiration ($E_{\sigma T}$), rainfall interception loss (E_i), total evapotranspiration (ET), reference evaporation (ET_0) (Allen et al., 1998), surface runoff (Q_s) and groundwater outflow (Q_g).

Date	P [mm]	n [day]	E_s [mm]	$E_{\sigma T}$ [mm]	E_i [mm]	ET [mm]	ET_0 [mm]	Q_s [mm]	Q_g [mm]
Mar 07	199	2	17	29	37	66	38	24	48
Apr 07	372	0	35	66	72	138	83	68	0
May 07	315	3	32	73	67	141	82	61	149
Jun 07	190	6	38	90	45	135	87	45	132
Jul 07	244	9	44	106	52	158	97	39	57
Aug 07	202	7	39	100	46	146	94	34	45
Sep 07	96	8	41	71	26	97	101	21	115
Oct 07	131	13	44	64	32	97	99	21	55
Nov 07	124	14	43	53	25	78	94	13	51
Dec 07	265	7	41	72	55	127	84	31	0
Jan 08	290	7	41	64	60	124	85	51	82
Feb 08	464	3	40	74	80	155	99	52	227
Mar 08	162	1	20	34	35	69	45	26	111
Total	3054	80	475	899	633	1532	1090	485	1071

amounted to 987 mm, exceeding the reference evaporation for that period (573 mm) by 414 mm.

Evapotranspiration

The average albedo (α) of the Campina forest at noon under dry conditions was 0.142 ($\sigma = 0.020$, $n = 283$) and decreased significantly ($p < 0.01$) to 0.108 ($\sigma = 0.039$, $n = 20$) when the canopy was wet. The campina forest albedo is significantly higher ($p < 0.01$) than that of the tall rainforest in the Asu catchment ($\alpha = 0.114$, $SD = 0.008$, $n = 295$), which would result in a lower net radiation. The daytime net radiation input at the Campina forest between 8:00 h and 18:00 h averaged 12.1 MJ day^{-1} ($\sigma = 3.7 \text{ MJ}$, $n = 210$), which was indeed significantly lower by 7% ($p < 0.05$) than the corresponding input for tall rainforest in the Asu catchment (13.0 MJ day^{-1} , $\sigma = 4.0 \text{ MJ}$, $n = 210$).

Time series of daily transpiration (E_s) and actual evapotranspiration ($ET = E_{\sigma T} + E_i$) are shown in Figure 4.1. Actual evapotranspiration ET amounted to 1532 mm (4.2 mm day^{-1} ; Table 4.1). This was similar to the potential evaporation (Penman open water evaporation; Penman, 1948), which amounted to 1503 mm or 4.1 mm day^{-1} , but was 442 mm higher than the FAO reference evaporation. Rainfall interception loss constituted 41% of total ET at 1.7 mm day^{-1} and dry canopy evaporation $E_{\sigma T}$ amounted to 2.5 mm day^{-1} . The ratio between the dry evapotranspiration and the FAO reference evaporation ($ET_0 = 3.0 \text{ mm day}^{-1}$) was 0.82.

Transpiration rates from the sapflow measurements E_s were 1.3 mm day^{-1} and therefore considerably lower than the dry evapotranspiration $E_{\sigma T}$. The lower value may be due to the selection of trees for the sapflow measurement, which were in the open campina forest, *i.e.* the area where trees were most stunted. Furthermore, sapflow measurements do not include soil evaporation, which may not have been negligible in the open campina forest, especially during periods with high surface soil moisture availability. The higher than normal rainfall in this year may also have suppressed transpiration as the canopy was wet for extended periods during daytime. The leaf wetness sensor indicated an average surface wetness of 13.7 h day^{-1} , and a daytime (8:00–18:00 h) canopy wetness of 3.5 h day^{-1} , or 35% of the time over a period of 330 days. However, this effect may partly have been compensated by lack of drought stress as the soil below a depth of 0.1 m remained above field capacity for the whole study period (Section 4.4.1). The dry evapotranspiration measured by the meteorological set-up is also likely to have been representative for a larger area, including the surrounding campinarana forest. As such the sapflow transpiration measurements may be viewed as a minimum value, representative for low-statured campina forest. If soil evaporation could be considered negligible, annual evapotranspiration based on sapflow and rainfall interception measurements would have amounted to 1108 mm, or 3.0 mm day^{-1} for the low-stature open campina forest. This would constitute 72% of ET_0 .

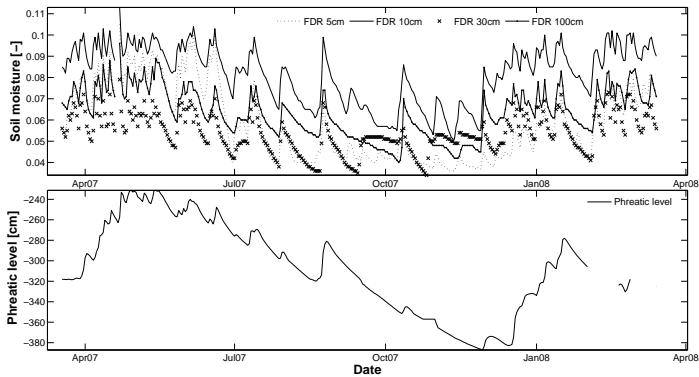


Figure 4.2: Time series of daily average soil moisture measured at four different depths in the soil of the campina forest.

Surface runoff

Stream runoff was low at 485 mm (Table 4.1), with daily values varying between 0.1 and 4.1 mm day⁻¹ (Figure 4.1). Stream runoff seemed to be dominated by groundwater contributions. Observed stormflow events were of very short duration and ended within hours after rainfall ceased. This is in line with the relative flatness of the area, the sandy – highly permeable – soil and, the relatively deep groundwater table (2–4 m below the surface) favouring infiltration and preventing the generation of overland flow even during the most intensive rainfall events. Stream discharge during the dry season (June–November) amounted to 173 mm, or 36% of the annual total listed in Table 4.1.

Soil moisture and groundwater

Soil moisture content (θ) varied from 0.036–0.105 at 0.05 m depth and from 0.040–0.096 at 1 m depth, with averages of 0.058, 0.082, 0.054 and 0.063 at 0.05, 0.10, 0.30 and 1.0 m depth, respectively. These averages are close to or above the moisture contents at field capacity θ_{FC} for the various depths, *i.e.* θ_{FC} = 0.082, 0.025 and 0.028 for the layers 0.0–0.25 m, 0.25–0.60 m and 0.60–1.20 m, respectively. In the study period, the 0.05 m layer remained above field capacity for 12% of the time, that at 0.1 m for 57%, and the deeper layers for 100% of the time.

The slightly higher θ at 0.1 m depth was in line with the somewhat finer soil texture (porosity of 0.48 at 0–0.25 m *versus* 0.37–0.40 for the deeper layers), perhaps due to some accumulation of organic matter in the surface layer. Time series of the daily averages at the four depths are presented in Figure 4.2. In spite of the many peaks in response to rainfall, a seasonal pattern can be distinguished with lower θ values from August until December, corresponding

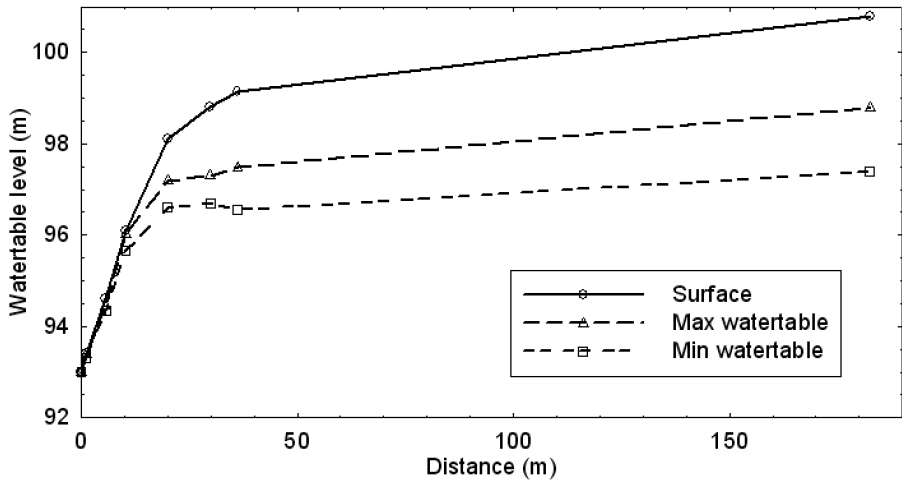


Figure 4.3: Range of the phreatic water levels obtained from manual measurements in the piezometers along a transect in the *Campina* catchment during the study period. The ground surface is represented by the solid line, whereas maximum and minimum phreatic levels are represented by the dashed lines. The symbols represent the locations of the piezometers where the manual measurements were done, with the left symbol representing the stream and the right symbol the tower piezometer.

with the recession observed in the stream hydrograph (Figure 4.1).

Phreatic water levels varied considerably during the study period, with rapid responses to rainfall events and gradual decreases during drier periods (Fig. 4.3). The largest variation in phreatic level (from 2.31 m until 3.87 m below the surface) was observed on the plateau near the micro-meteorological tower at 182 m distance from the stream. The average water level depth at that location was 3.06 m below the surface. The regular manual water level measurements, which did not capture the full variation, made in the other piezometers indicated that the phreatic levels decreased gradually with a low gradient of $0.009(\pm 0.001)$ from the tower piezometer on the plateau to PC6 at the edge of the plateau, and then with a much steeper gradient of $0.127(\pm 0.006)$ to PC1 at 1 m distance from the stream. This sudden change in gradient may be caused by a decrease in the hydraulic conductivity (K_s) due to an increasing clay content in the soil near the valley.

To estimate groundwater drainage from the catchment (Table 4.1), monthly water balances were calculated. Changes in soil moisture storage were assessed from the soil moisture profile data. Groundwater storage changes were estimated from changes in the phreatic level in the piezometer on the plateau near the meteorological tower using a specific yield of 25% appropriate for medium sands (Johnson, 1966). Monthly changes in the soil moisture storage

ranged between -58 mm in June 2007 and 65 mm in February 2008. The soil moisture storage change over the whole study period was negligible at -3 mm. Storage changes in groundwater showed a larger range, from -115 mm in the dry month of September 2007 to 190 mm in the April 2007. Again, the change in groundwater storage was small over the whole period at -17 mm. As such, total storage change over the study period amounted to -20 mm. Incorporating these storage changes (ΔS) in the monthly water balances provided the monthly estimates of groundwater drainage ($Q_g = P - ET - Q_s - \Delta S$) as listed in Table 4.1. Groundwater discharge during the dry season (June–November) amounted to 454 mm, or 42% of the annual total listed in Table 4.1. Groundwater drainage was much larger than stream discharge, accounting for 69% of total discharge (1557 mm, or 4.3 mm day⁻¹).

4.4.2 Campina forest hydrochemistry

Dissolved organic and inorganic carbon

The soil moisture in the unsaturated zone showed a wide range of DOC concentrations. As mentioned earlier, soil moisture samples could not always be collected from all samplers and the bulk concentrations were therefore also influenced by the difference in volumes that could be extracted from the three samplers under bare soil, forest and intermediate positions. The results are therefore presented in Table 4.2 as medians and ranges for the surface (0.0–0.4 m) and subsurface (0.4–0.6 m) layers.

The pH averaged 4.14(± 0.37), 4.02(± 0.13) and 3.88(± 0.19) for soil moisture, groundwater and stream flow, respectively. The pH of stream water was significantly lower than that of groundwater (at $p < 0.05$). The pH of soil moisture was not significantly different from that of groundwater or river water. At these low pH values DIC can be considered to be entirely in the form of dissolved CO₂ (Appelo and Postma, 2005; Sousa et al., 2008). The pH of stream water was positively correlated to the DIC concentration, whereas an inverse trend was observed for DOC.

DIC concentrations were invariably low in soil moisture, ranging from 0.24–1.04 mg l⁻¹. Intermediate values were observed in the stream (0.63–2.23 mg l⁻¹), whereas the highest values were for groundwater (0.69–11.41 mg l⁻¹). Average groundwater DIC concentration was significantly higher than those in the stream or of soil moisture ($p < 0.01$), whereas that of soil moisture was significantly lower than that of stream flow ($p < 0.01$). Groundwater and stream flow DIC concentrations both showed seasonal patterns, with the highest values observed in the dry period (June – December). Changes in DOC and DIC in groundwater along a transect from the stream to the plateau are shown in Figure 4.4. The mean DIC concentration in groundwater reached a maximum of 6.5 \pm 3.2 mg l⁻¹ (n= 11) in the valley at 6 m from the stream where the soil was fine textured and poorly-drained, and was somewhat lower at 4.2 \pm 1.0 mg l⁻¹ (n= 8) in the sandy soil in the open Campina forest at 182

Table 4.2: Statistics of dissolved inorganic (DIC) and organic (DOC) carbon concentrations in soil moisture (SM) at different depths, groundwater (all piezometers) and stream water obtained from monthly samples between February 2007 and March 2008. The number of samples is represented by n.

	DIC				DOC			
	Average [mg l ⁻¹]	St.dev. [mg l ⁻¹]	Median [mg l ⁻¹]	n	Average [mg l ⁻¹]	St.dev. [mg l ⁻¹]	Median [mg l ⁻¹]	n
SM _{0.0-0.4}	0.51	0.20	0.42	20	35.8	19.4	37.4	24
SM _{0.4-0.6}	0.98	1.40	0.35	5	9.0	4.5	7.1	8
SM _{0.0-0.6}	0.61	0.63	0.42	25	29.1	20.6	24.0	32
Groundwater	5.07	2.58	4.47	83	29.0	6.0	27.4	73
Stream flow	1.54	0.56	1.54	11	30.8	3.9	30.7	11

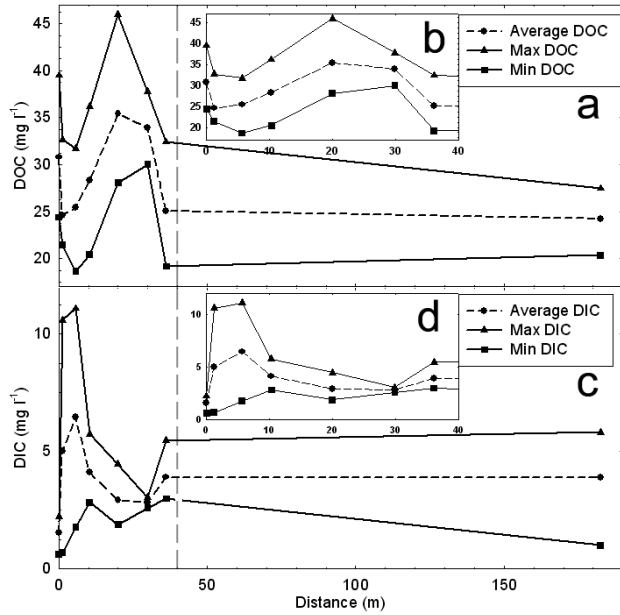


Figure 4.4: (a) DOC concentrations in groundwater at varying distances from the stream. The dashed line with dark circles represent the average DOC concentrations throughout the period; the continuous lines with dark triangle and dark square represent the maximum and the minimum concentration observed, respectively. (b) Close up of the left side of the gray dashed line in figure (a) keeping the same units. (c) and (d) show has the same structure as (a) and (b) presenting DIC concentration in groundwater *versus* distance from the stream.

m from the stream (Figure 4.4).

The contrast between the DIC concentrations in the valley groundwater close to the stream (6.5 mg l^{-1}) and in stream water (1.5 mg l^{-1}) suggests that rapid evasion of CO_2 occurred from when exfiltrating groundwater entered the stream, significantly lowering the DIC concentration downstream at the river water sampling point by about 5 mg l^{-1} . Linear regression yielded the following relation between DIC concentration in stream water (DIC_s in mg l^{-1}) and the C evasion concentration (C_e in mg l^{-1}), *i.e.* the difference in DIC concentrations in groundwater at 6 m from the stream and in stream water ($R^2 = 0.37$):

$$C_e = 3.09 \cdot \text{DIC}_s + 0.15 \quad (4.1)$$

DOC concentrations in the unsaturated top 0.4 m of the soil showed a very wide range of $0.7\text{--}70.7 \text{ mg l}^{-1}$, reflecting different volume contributions from bare soil (low DOC) and below-tree (high DOC) soil moisture to the bulk samples. Water could only be extracted under very wet conditions from slightly coarser sand layer at 0.5 and 0.6 m depths. This may explain the much lower DOC values, and slightly higher DIC concentrations observed for these depths. The average DOC concentrations in soil moisture, ground- or river water were not significantly different. Groundwater did not show a clear seasonal pattern in DOC concentrations, in contrast to DOC in the stream, where concentrations reached a maximum of 39.5 mg l^{-1} in April 2007 and gradually decreased to a minimum of 24.4 mg l^{-1} at the end of the dry period (November 2007). Both DOC and DIC showed variations in concentrations along the slope towards the stream. This coincided with a change in soil texture and presumably also in litter decomposition processes due to the near-permanent saturation of the valley soil (Figure 4.4).

4.4.3 Inorganic chemistry

Overviews of inorganic ion concentrations and electrical conductivity (EC) observed in the Campina catchment and in the Asu rainforest catchment are presented in Tables 4.3 and 4.4.

The observed concentrations were all very low, and were sometimes below detection limits. The average concentrations and EC in soil moisture were significantly higher ($p < 0.05$) than those in groundwater or stream water except for NH_4^+ and Ca^{2+} , which did not show significant differences between water types. The K^+ concentration in groundwater was significantly higher than in stream water ($p < 0.05$). Between groundwater and stream flow, significant differences ($p < 0.05$) were observed for Na^+ , K^+ , Cl^- , NO_3^- and PO_4^{3-} concentrations, which were all higher in the former. Expressed in meq l^{-1} , the dominant cations were Na^+ and Ca^{2+} and dominant anions were Cl^- and SO_4^{2-} .

The EC of groundwater in the Campina catchment was significantly higher ($p < 0.05$) than that in both the riparian zone or campinarana – ecotone forest

Table 4.3: Averages and standard deviations of cation concentrations in Campina catchment soil moisture (SM, n= 13), groundwater (all piezometers, n= 71) and stream water (n= 11) obtained from monthly samples between February 2007 and March 2008. Averages and standard deviations for groundwater in the riparian forest (n= 7) and campinarana – ecotone forest (n= 12) in the Asu catchment are provided for comparison. The < sign indicates that one or more samples had concentrations below detection limits, the detection limit was then used to calculate averages and standard deviations.

	Na ⁺ [mg l ⁻¹]	K ⁺ [mg l ⁻¹]	Ca ²⁺ [mg l ⁻¹]	Mg ²⁺ [mg l ⁻¹]	NH ₄ ⁺ [mg l ⁻¹]
Campina catchment					
SM _{0.0–0.6}	1.45±0.61	0.26±0.17	0.15±0.15	0.08±0.05	0.06±0.08
Groundwater	0.55±0.39	0.19±0.14	<0.10±0.07	<0.02±0.02	0.05±0.05
Stream flow	0.37±0.26	0.12±0.04	0.09±0.12	0.01±0.01	0.07±0.17
Asu rainforest catchment (groundwater)					
Riparian zone	0.77±0.92	0.28±0.32	0.11±0.07	0.03±0.02	0.04±0.02
Campinarana	1.07±0.92	0.31±0.17	0.09±0.07	0.02±0.01	0.03±0.03

in the Asu rainforest catchment, which could be attributed to the difference in DOC concentrations. In the riparian zone in the Asu catchment, the Mg²⁺ concentration was significantly higher than in the groundwater in the Campina catchment, whereas the SO₄²⁻ concentration was lower in the campinarana – ecotone forest (Tables 4.3 and 4.4). The concentrations of Cl⁻ in the Asu rainforest groundwater were consistently higher than that in the Campina catchment, which arguably could be due to higher evapotranspiration in the former. The other ion concentrations in groundwater showed no significant differences between sites and the pH was only varied within a narrow range of 4.01–4.13.

Carbon and nutrient exports

The contribution of inorganic ions to the EC ($\mu\text{S cm}^{-1}$) of water can be estimated as $EC = 50 \cdot (\sum \text{anions} + \sum \text{cations})$ if the sum of cations and anions is below 6 meq l^{-1} (Appelo and Postma, 2005). This would yield $EC_{\text{inorganic}}$ values between 2.2 and $7.2 \mu\text{S cm}^{-1}$ for stream water, which is much lower than the observed conductivity values (Table 4.4). This suggests that the EC was highly influenced by DOC concentrations. As the pH was well below 4.5, DIC would have been in the form of dissolved CO₂ and would therefore not contribute to the EC. Linear regression analysis of EC and corresponding DOC and DIC concentrations resulted in the following equations:

$$\text{DOC} = 0.639 \cdot \text{EC} + 3.984 \quad (4.2)$$

$$\text{DIC} = -0.105 \cdot \text{EC} + 5.956 \quad (4.3)$$

Table 4.4: Averages and standard deviations of *EC* and inorganic anion concentrations in soil moisture (SM, n=13), groundwater (all piezometers, n=71) and stream water (n= 11) obtained from monthly samples between February 2007 and March 2008. Averages and standard deviations for groundwater in the riparian forest (n= 7) and campinarana – ecotone forest (n= 12) in the Asu catchment are provided for comparison. The < sign indicates that one or more samples had concentrations below detection limits, the detection limit was then used to calculate averages and standard deviations.

	EC [$\mu\text{S cm}^{-1}$]	Cl ⁻ [mg l ⁻¹]	SO ₄ ²⁻ [mg l ⁻¹]	NO ₃ ⁻ [mg l ⁻¹]	PO ₄ ³⁻ [mg l ⁻¹]
Campina catchment					
SM _{0.0-0.6}	58.6±20.4	1.25±0.66	0.41±0.28	0.16±0.17	<0.016±0.018
Groundwater	39.4±6.6	0.35±0.26	<0.06±0.06	<0.02±0.03	<0.006±0.005
Stream flow	42.0±4.5	0.22±0.17	<0.05±0.10	<0.01±0.01	<0.005±0.000
Asu rainforest catchment (groundwater)					
Riparian zone	30.0±6.5	0.45±0.12	<0.04±0.06	<0.02±0.01	<0.006±0.001
Campinarana	31.0±14.9	0.75±0.53	<0.03±0.03	<0.03±0.03	<0.009±0.008

Table 4.5: Exports of DOC and DIC in stream water (Q_s) groundwater (Q_g) and in total (C_{export}) for a 366 day period between March 2007 and March 2008. An estimate of C evasion, based on difference in DIC concentrations in valley groundwater and river water, is given for comparison.

Date	Q_s -DOC [g m ²]	Q_s -DIC [g m ²]	Q_g -DOC [g m ²]	Q_g -DIC [g m ²]	C_{export} [g m ²]	C evasion [g m ²]
Mar 07	0.79	0.03	1.35	0.20	2.37	0.09
Apr 07	2.33	0.06	0.00	0.00	2.39	0.21
May 07	2.07	0.06	4.84	0.71	7.69	0.20
Jun 07	1.42	0.06	4.06	0.80	6.34	0.20
Jul 07	1.15	0.07	1.80	0.54	3.56	0.22
Aug 07	0.95	0.07	1.37	0.29	2.68	0.21
Sep 07	0.60	0.04	3.42	0.41	4.47	0.13
Oct 07	0.60	0.04	1.56	0.43	2.63	0.12
Nov 07	0.36	0.03	1.43	0.21	2.02	0.09
Dec 07	0.91	0.06	0.00	0.00	0.97	0.18
Jan 08	1.69	0.06	2.72	0.44	4.91	0.19
Feb 08	1.61	0.08	6.08	1.01	8.77	0.24
Mar 08	0.81	0.04	3.05	0.48	4.38	0.12
Total	15.3	0.69	31.7	5.5	53.2	2.2

yielding coefficients of determination of 0.54 and 0.72, respectively ($n=11$). Note that the DIC concentration was negatively correlated with EC, whereas DOC showed a positive correlation. These equations were subsequently used to estimate daily DOC and DIC concentrations for stream water based on the continuously measured EC values and these were combined with the daily surface runoff totals to obtain DIC and DOC exports. An estimate of the C evasion from the stream was obtained using Equation 4.1 to calculate evaded DIC, in combination with the daily surface runoff totals.

Groundwater exports were calculated using the average concentrations of the monthly samples from all piezometers. The carbon exports obtained from combining stream discharge and groundwater drainage with corresponding daily and monthly DOC and DIC concentrations are presented in Table 4.5.

The total DOC export in surface runoff amounted to 15.3 g m⁻² over the 366-day study period, whereas the corresponding DIC export was much lower at 0.7 g m⁻² (4%). During the dry season DOC and DIC exports accounted for 33% and 44% of total exports in stream discharge, respectively. Exports of carbon were much larger in groundwater drainage, accounting for 70% of the total annual export. The difference in exports between seasons was less pronounced for groundwater drainage, with values of 43% and 49% of annual DOC and DIC exports occurring during the dry season, respectively. C-export during the dry season accounted for 41% of the annual total. Based on the

mean DOC concentration of 1.32 mg l^{-1} in rainfall from observations in the Anavilhanas Archipelago $\sim 90 \text{ km}$ to the West (Filoso et al., 1999) a DOC input with rainfall of 4.0 g m^{-2} would be obtained, resulting in a net hydrological DOC export of 49.2 g m^{-2} from the Campina catchment. The estimated evasion of C-CO₂ was about three times the DIC export in surface runoff, but was only 4% of the hydrologic C export.

To calculate macro-nutrient and other inorganic ion exports, the concentrations obtained from the monthly samples were used in combination with the monthly river and groundwater discharge values (Table 4.1). Dissolved inorganic nitrogen (DIN) exports were calculated as the sum of N-NO₃ and N-NH₄ exports, as NO₂ concentrations remained below detection limits. The results are shown in Table 4.6. For comparison, nutrient inputs in rainfall were also calculated using the concentrations in rainfall observed in the Lake Calado study ($\sim 70 \text{ km}$ to the SW) by Lesack and Melack (1991) and in the Anavilhanas Archipelago ($\sim 90 \text{ km}$ to the SW) by Filoso et al. (1999), in combination with annual rainfall measured during the present study.

With the exception of N – because of the lower NH₄⁺ concentration in groundwater – all inorganic ion exports in groundwater were higher than those in surface water. A comparison between the values presented in Tables 4.5 and 4.6 shows that annual inorganic ion exports were only a small fraction of those of carbon. A comparison of estimated inputs via rainfall and exports in water draining from the catchment suggested small net gains for N, P, Mg and S, a near-balance for Ca and DIP, and small losses for the other ions. Rainfall H⁺ input was estimated at 0.027 g m^{-2} , whereas exports in the acidic surface (0.074 g m^{-2}) and groundwater (0.103 g m^{-2}) were much larger, resulting in a net loss for H⁺ of 0.15 g m^{-2} .

4.5 Discussion

4.5.1 Water balance aspects

The catchment water balance could not be used to quantify evapotranspiration because of uncertainties related to undetected groundwater outflow from the flat sandy area. Rainfall, evapotranspiration, surface runoff and soil and groundwater storage were therefore measured independently to allow quantification of groundwater drainage. The study was done in a wet period with annual rainfall (3054 mm) being over 800 mm above the long-term annual average observed at Manaus (2211 mm, 1927–2007; Satyamurty et al., 2010).

Evapotranspiration was determined separately for wet and dry canopy conditions. Sapflow based transpiration of the low-statured campina forest was low at 475 mm year^{-1} , or 1.3 mm day^{-1} . This was about 60% of the transpiration rate of *terra firme* lowland rainforest in the nearby Asu catchment as modelled by Broedel (2012, ; 766 mm year^{-1} in 2007–2008) using the CLM model (Dai et al., 2003) calibrated on long-term records of soil moisture and

Table 4.6: Exports of macro-nutrients dissolved inorganic N (DIN; $\text{NO}_3\text{-N} + \text{NH}_4\text{-N}$) and P (DIP; $\text{PO}_4\text{-P}$), K, Ca, Mg and other major inorganic ions (S as $\text{SO}_4\text{-S}$) in stream water (Q_s), groundwater (Q_g) and the total export (Q_t) for the 366 day period between March 2007 and March 2008. All values are reported in g m^{-2} and the $<$ sign indicates that the detection limit was used if one or more samples had concentrations below detection limits. Estimates of nutrient inputs with rainfall using observations from the Lake Calado study by (P_{LC} ; Lesack and Melack, 1991) are given for comparison and were used to calculate net exports.

	DIN	DIP	K	Ca	Mg	Na	Cl	S
P_{LC}	0.356	0.004	0.065	0.162	0.074	0.127	0.349	0.330
Q_s	<0.042	<0.001	0.064	0.058	<0.007	0.205	0.119	0.005
Q_g	<0.038	<0.002	0.217	<0.106	<0.016	0.630	0.290	<0.020
Q_t	<0.080	<0.003	0.281	<0.164	<0.024	0.835	0.509	<0.024
$P_{LC} - Q_t$	0.276	0.001	-0.216	-0.002	0.051	-0.708	-0.160	0.306

eddy covariance data. Heath forest transpiration is difficult to measure and reliable annual estimates for other locations are not available yet. The presently observed rate is similar to that given by Becker (1996b) for lowland heath forest in Brunei (1.2 mm day^{-1}) in dry and wet periods lasting several months. In his sapflow study, Becker (1996b) observed no differences between tall dipterocarp rainforest and heath forest. A much lower average transpiration value of 0.7 mm day^{-1} was reported during a three-month dry to wet season transition period for pre-montane non-water stressed heath forest on sandy soils in the *Selva Alta* of Peru at 1400 m a.s.l. (Mette and Zimmermann, 2003). In the same study transpiration from adjacent rainforest on sandy soils was also low, albeit 38% higher than that of the heath forest, at 1.0 mm day^{-1} . Transpiration rates did not show any response to the dry season in both pre-montane forests (Mette and Zimmermann, 2003).

At 899 mm year^{-1} , dry canopy evaporation as determined from the temperature variance energy balance method was at the lower end of the range ($885\text{--}1285 \text{ mm year}^{-1}$) obtained from a review of a large number of tall tropical rainforest studies by Bruijnzeel (1990). Because of the sclerophyllous nature of the leaves of campina trees, the low LAI (2.4) and the results from sapflow measurements, a lower rate would be expected. It should be realised though that the micro-meteorological set-up provided values for a larger area than that of the low-statured campina forest in which the tower was located, incorporating contributions from the surrounding taller campinarana forest.

Annual evapotranspiration based on sapflow and rainfall interception (3.0 mm day^{-1}) for the low-stature open campina forest amounted to 83% of the nine-year average tall rainforest evapotranspiration ($3.6 \pm 0.3 \text{ mm day}^{-1}$) as presented by Broedel (2012) for the Asu rainforest catchment at about 19 km distance.

Interception loss for the campinarana and campina forests at the study site (16% of P_G) was in the range of that found in Kalimantan for tall heath forest (10% of P_G) and stunted heath forest (21% of P_G) by Vernimmen et al. (2007). The observed value is also in the range of that observed for the nearby tall rainforest in the Asu catchment by (13–23%; Cuartas et al., 2007) and lowland Amazonian rainforest elsewhere in Brazil and Colombia (Tobón Marin et al., 2000). This suggest that the differences in height, structure, LAI and tree composition between the campina, campinarana and tall lowland rainforests had little impact on wet canopy evaporation rates, which was also observed in the study of Vernimmen et al. (2007). However, as stemflow was not measured in the present study, the actual interception loss could well have been overestimated.

Evapotranspiration ($E_{\sigma T} + E_i = 1532 \text{ mm year}^{-1}$) for this site was at the higher end of the range observed for tropical rainforests ($1300\text{--}1500 \text{ mm year}^{-1}$; Bruijnzeel, 1990) and similar to that obtained from a 3-year catchment water balance by Tomasella et al. (2007) for the nearby tall *terra firme* forest Asu catchment ($1370\text{--}1470 \text{ mm year}^{-1}$). This would suggest that a change from the present campinarana - ecotone forests in the sandy poorly-drained

valleys in the area to a well-drained upland type of campina - campinarana forest would have little impact on the annual moisture supply to the atmosphere. However, the result of this study was for an exceptionally wet year in which rainfall interception loss accounted for 41% of the evapotranspiration loss and soil water stress did not play any role as soil moisture content in the profile remained at field capacity throughout the year. Furthermore, net radiation and evapotranspiration, and therefore the moisture supply to the atmosphere, were indeed lower than those for tall rainforest during dry periods. A such a change in forest cover to campina – campinarana forest is likely to cause a decrease in evapotranspiration during dry periods.

The combined surface and groundwater drainage from the Campina catchment was slightly higher (1557 mm; runoff coefficient of 0.51) than that observed in 6.6 km² Asu catchment (1362 mm; runoff coefficient of 0.46; Waterloo et al., 2006) for the wet year of 2002 ($P=2976$ mm). Leopoldo et al. (1995) also observed lower runoff coefficients, between 0.29 and 0.38, for the 1.3 km² Barro Branco rainforest catchment north of Manaus. This suggests a higher evapotranspiration loss in tall rainforest catchments than in the Campina forest catchment.

4.5.2 Biogeochemistry

The present work is the first attempt to give annual estimates of DIC and nutrient exports from a heath forest catchment in Amazonia. Richey et al. (1990) measured an overall mean DIC concentration of 9.4 mg l⁻¹ for the Amazon River at a pH between 6.5 and 7.2 and a DIC concentration for the Rio Negro (pH \approx 4) of less than 3 mg l⁻¹. DIC concentrations in groundwater in the campina forest were within this range at an average of 5 mg l⁻¹, but those in the stream draining the area were lower at 1.5 mg l⁻¹ at an average pH of 3.9 (Table 4.2). The mean DIC concentration in groundwater in the Campina catchment was significantly lower ($p < 0.05$, $\nu = 7$) than that in groundwater of the saturated valley bottom in the Asu catchment (5.1 ± 2.6 vs. 6.9 ± 1.8 mg l⁻¹; Zanchi et al., 2011), but compared well with that observed under the campinarana – ecotone valley forest (5.5 ± 2.7 mg l⁻¹), where the phreatic level was at ~ 1 m below the surface. Other studies in Amazonian rivers concur that DIC concentrations are relatively low (Gomez et al., 1991; Johnson et al., 2006; Sousa et al., 2008). Sousa et al. (2008) observed a similarly low mean DIC concentration of 2.0 ± 0.7 mg l⁻¹ in a stream draining a sandy rainforest catchment in Acre State (Brazil), albeit at a higher pH of 5.4. In the same study a higher DIC concentration of 4.8 ± 1.6 mg l⁻¹ was observed in a stream draining a silty rainforest catchment with a pH of 6.4. An average DIC concentration of 4.5 mg l⁻¹ was reported for outflow of the Amazon river into the Atlantic Ocean (Druffel et al., 2005). Sousa et al. (2008) also observed a similar seasonal variation in river DIC concentrations with the lowest DIC concentration (1.0 mg l⁻¹) observed in the wettest month. The lower DIC concentration in the Campina forest stream in comparison to that in groundwater

suggests rapid evasion of CO₂ from the stream into the atmosphere.

In contrast, the mean DOC concentration in groundwater in the Campina catchment was significantly higher ($p < 0.01$, $\nu = 6$) than that in the saturated valley bottom in the Asu catchment (29.0 ± 6.0 vs. 17.9 ± 4.7 mg l⁻¹; Zanchi et al., 2011), as well as in that under the campinarana – ecotone forest (16.4 mg l⁻¹; $p < 0.01$, $\nu = 12$). This suggests that a lowering of the groundwater level in the now permanently saturated area along the stream in the Asu rain-forest catchment could have a significant impact on the concentrations of both DIC and DOC in groundwater exfiltrating into the river, thereby affecting the the hydrological carbon export.

Finlay (2003) reported that smaller streams had lower DIC concentration than larger ones. This was attributed to higher CO₂ evasion in small streams, because of higher groundwater inputs (Sousa et al., 2008). A low DIC concentration and high pH would reduce the dissolved CO₂ (Johnson et al., 2006) to be released from the soil to the atmosphere or export to the stream and then release in form of CO₂ on the stream water. According to Zanchi et al. (2012), the total soil CO₂ emission was 3.08 ± 0.8 μmol m⁻² s⁻¹, which was similar to the other forest type for central Amazonia, perhaps this is a compensation of low DIC concentration, resulting in a high CO₂ emission for this total area.

An overview of carbon and nutrient exports for Amazonian forests is given in Table 4.7.

Total dissolved carbon export from the Campina catchment amounted to 53.2 g m⁻² year⁻¹, which is significantly larger than that observed for Rio Negro black water rivers draining lowland rainforest in central Amazonia (DOC exports of 9 – 23 g m⁻² year⁻¹, Table 4.7; Waterloo et al., 2006; Moreira-Turcq et al., 2003b; Richey et al., 1990) or for the more sediment-rich Purus River in southwestern Amazonia (15 g m⁻² year⁻¹; Salimon et al., 2012).

DOC export estimates were presented earlier for this site with runoff taken as 50% of mean annual rainfall at Ducke Reserve in Manaus (2250 mm; McClain et al., 1997). DOC concentrations in both stream- and groundwater reported by McClain et al. (1997, 38.4 ± 6.6 and 35.9 ± 9.9 mg l⁻¹, respectively) were significantly higher than those observed in this study (30.8 ± 3.9 and 29.0 ± 6.0 mg l⁻¹, respectively). The calculated DOC export is somewhat lower than that observed in this study because of the lower drainage estimate of McClain et al. (1997). The DOC export is much higher than that observed by Waterloo et al. (2006) in the Asu catchment, where soils in the valley are of similar texture but remain permanently near saturation. This would suggest that a permanent lowering of the phreatic level in the valleys due to climate change could cause a doubling in DOC concentrations from the 15 mg l⁻¹ observed in the Asu catchment (Waterloo et al., 2006) to 31 mg l⁻¹ measured in campina forest stream water. Such an increase should not be expected in the hydrological export of DOC as river runoff would certainly decrease much under the drier conditions.

In a study comparing flowpaths of dissolved carbon in paired Oxisol and Ultisol catchments in Mato Grosso, SW Amazon, Johnson et al. (2006) re-

Table 4.7: Carbon and dissolved inorganic nutrient exports obtained in the present study (Present) and an earlier studies at this site (Campina; McClain et al., 1997) and in other central Amazonian catchments such as Lake Calado (Lesack, 1993), Asu (Waterloo et al., 2006), Ducke (Brinkmann, 1983) and Barro Branco (Franken and Leopoldo, 1984). DIN = dissolved inorganic nitrogen, DON= dissolved organic nitrogen and TN= total nitrogen.

Species	Present [g m ⁻² year ⁻¹]	Campina [g m ⁻² year ⁻¹]	Lake Calado [g m ⁻² year ⁻¹]	Asu [g m ⁻² year ⁻¹]	Ducke [g m ⁻² year ⁻¹]	B. Branco [g m ⁻² year ⁻¹]
DIC	6.2	-	-	-	-	-
DOC	47.0	40.0	-	9.3–22.7	-	-
DIN	<0.080	0.05	0.286	-	0.36	-
DON	-	0.80	0.076	-	2.56	-
TN	-	0.85	0.431	-	2.90	-
DIP	<0.003	-	0.001	-	0.03	<0.001
TP	-	-	0.008	-	-	-
K	0.281	-	0.047	-	-	0.039–0.090
Ca	<0.164	-	0.054	-	0.09	-
Mg	<0.024	-	0.033	-	0.05	-
Na	0.835	-	0.260	-	-	0.090–0.191
Cl	0.509	-	0.783	-	-	-
S	<0.024	-	0.107	-	-	0.491–1.068
H	0.177	-	0.040	-	-	-
Fe	-	-	-	-	0.36	-

ported large dissolved organic C-fluxes through rapid flowpaths after high rainfall in the catchment, and high biogenic CO₂ fluxes reaching groundwater through slow flowpaths that allowed sorption and decomposition to occur. In these soils with a much higher sorption capacity than the sandy campina soil, the DOC concentration decreased with depth in the profile, whereas CO₂ increased. Such a decrease in DOC concentration was not observed in the campina forest soil. It is likely that rapid transport of litter layer and root leachates in water travelling through the coarse sandy unsaturated zone could occur after larger rainfall events, which may explain the high DOC concentrations observed in groundwater, as the soil sorption capacity was very low and the residence time in the unsaturated zone would be short, leaving little time for decomposition to occur. In the Asu rainforest catchment, the highest DOC concentrations were observed in stormflow (Waterloo et al., 2006). In the broad sandy valleys at Asu, saturation overland flow in response to rainfall provides a fast pathway for water containing a high concentration of DOC from litter layer and root leachates to the river. A future climate with less rainfall would make the mechanism of saturation overland flow contributing to hydrological carbon export less important, which could be partly compensated by an increase in DOC in valley groundwater. Differences between inorganic ion concentrations in groundwater in the Campina catchment and in the valley of the Asu catchment were not significant, with exceptions for Mg²⁺, SO₄²⁻ and Cl⁻. As such a lowering of the groundwater level in the valley of the Asu catchment may not lead to changes in the inorganic ion composition, but would reduce nutrient exports from the catchment due to the a decrease in runoff.

Herrera et al. (1978) and Brinkmann (1985) reported that strongly weathered soils are generally characterized as having relatively conservative nutrient cycles that result in low losses of solutes to streams or groundwater. Exports of nutrients are invariably low in the central Amazonian rainforest area (Table 4.7), where Ultisol, Oxisol and Spodosol soils were strongly leached. The low export of macro-nutrients in this study is therefore not surprising. Comparison of Dissolved Inorganic N (DIN) between Campina and rainforest sites indicates that DIN exports are much lower from the former (Table 4.7). However, total N export from the Campina forest seems to be higher than that of terra firme rainforest because of higher Dissolved Organic N (DON) exports (McClain et al., 1997; Lesack, 1993; Brinkmann, 1983).

Potassium, sodium and calcium exports from the Campina catchment also seem to be higher than those observed for Amazonian terra firme rainforest, whereas that of sulphur seems lower (Table 4.7). This suggests that the campina – campinarana forest could be less efficient in retaining these nutrients than the terra firme rainforest. It is not known if this is due to the coarse sandy soil with low nutrient retention capacity and favouring rapid transport of nutrients to groundwater, or to the demands of the vegetation for specific nutrients to sustain its maintenance and slow growth.

4.6 Conclusions and recommendations

This study aimed to provide insight into possible changes in evapotranspiration, soil water chemistry, carbon and inorganic nutrient exports from the valleys in central Amazonian rainforest catchments that could occur in response to a drier future climate. This was done by study of a campina – campinarana heath forest on leached white sandy quartz soil as a lowering of phreatic water levels in the poorly-drained sandy valleys of rainforest catchments could change soil and drainage conditions in these valleys such that these become similar to those in upland campina – campinarana heath forest.

Quantification of the water balance components of the Campina catchment showed that sapflow transpiration rates in open campina forest were very low at 1.3 mm day^{-1} . Micrometeorological estimates of dry evaporation from the campina – campinarana forest were on the lower end of those observed for tropical rainforest at 2.5 mm day^{-1} . Rainfall interception loss accounted for 16% of gross precipitation and brought the actual evaporation (4.2 mm day^{-1}) in this exceptionally wet year of study to a value on the higher end of the range published for tropical rainforest. However, in normal years, when the rainfall interception loss is lower and soil moisture stress may play a role, evapotranspiration may be expected to be lower than that of rainforest growing on soils with higher plant water availability. A comparison of net radiation at tall rainforest and campina forest indicated that 7% less energy would be available in the latter due to the higher albedo (0.14 *versus* 0.11) and presumably higher surface temperatures. The higher runoff coefficient for the Campina catchment also indicates a lower evaporation in comparison to that in rainforest.

As yet, no information is available about the evapotranspiration rate of campinarana - ecotone forest in poorly-drained valleys in central Amazonia. It is therefore not possible to state if a transition of such valley forests to a drier campinarana - campina forest in response to climate change would alter catchment evapotranspiration. This study showed that there is a need for long-term evapotranspiration measurements in upland campina – campinarana and valley campinarana - ecotone forests to assess the impact of water stress in the former and to determine current evapotranspiration rates in the latter. In addition, comprehensive rainfall interception studies, including stem flow measurements, are needed in Amazonian campinarana – ecotone valley forest and in campinarana – campina heath forests to extend our current knowledge. This would allow better predictions on the impacts of climate change on the spatial variation of evapotranspiration in rainforest catchments forests.

In this very permeable catchment, surface runoff was predominantly from exfiltration of groundwater, and only amounted to 31% of total drainage. DIC concentrations were significantly lower in surface runoff than in groundwater drainage, which would suggest evasion of CO_2 from exfiltrating groundwater in the source area. DOC concentrations were much higher and dissolved carbon exports were therefore dominated by DOC exports. DIC concentrations in groundwater in the Campina catchment were lower than those in the valley

of the Asu rainforest catchment, whereas DOC values were much higher. This suggests that a drier climate with falling phreatic levels in the valleys would result in higher DOC and lower DIC concentrations in groundwater, and therefore also in the baseflow of these rainforest catchments. Hydrological carbon export from the Campina catchment was much higher than that from the Asu rainforest catchment. As runoff from rainforest catchments is likely to decline in a drier climate, it is questionable if higher DOC concentrations would cause an increase in carbon exports. Further studies are needed to understanding the processes that lead to DOC formation in the well-drained campina forest soils and in the poorly drained valley soils of rainforest catchments.

Concentrations of inorganic nutrients in groundwater were similar in the Campina catchment and in the riparian zone in the Asu rainforest catchment. The export of nutrients from the Campina catchment was low, but exports of Na, K and Ca were higher than those observed in rainforest catchments, whereas that of S was lower. The nutrient study suggested that the Campina catchment was somewhat less efficient in retaining K and Ca, which could be arguably related to the low nutrient retention capacity of the sandy soil. The impact of a lowering of the phreatic level on the inorganic ion concentrations in groundwater in the valleys of rainforest catchments can therefore be viewed as minimal.

Chapter 5

Influence of drainage status on soil and water chemistry, litter decomposition and soil respiration in central Amazonian forests on sandy soils ⁴

5.1 Introduction

The Amazon region (6.6 million km²) contains more than half of Earth's remaining tropical rain forest (Houghton et al., 1995), which accounts for 30–50% of the total global primary production (Dixon et al., 1994). Over the past decades about 750.000 km² of the Brazilian Amazon has been deforested (Nobre and Borma, 2009) which, 11.968 km² y⁻¹ between 2007 and 2008 (INPE, 2008). The large-scale conversion of rain forest to agricultural land may cause changes in local, as well as global water and carbon cycles (Taylor and Lloyd, 1992). Changes in the hydrological cycle may potentially lead to dangerous positive feedback effects in the carbon cycle through tree mortality (drought), changes in CO₂ emissions and in the forest CO₂ sink strength, de-

⁴The contents of this chapter have been published as F. B. Zanchi, M. J. Waterloo, A. J. Dolman, M. Groenendijk, J. Kesselmeier, B. Kruijt, M. A. Bolson, F. J. Luizão, A. O. Manzi. Influence of drainage status on soil and water chemistry, litter decomposition and soil respiration in central Amazonian forests on sandy soils. *Ambi-Agua*, Taubat, v. 6, n. 1, p. 6-29, 2011. (doi:10.4136/ambi-agua.170)

creased biomass production and reduced evapotranspiration by the remaining forest. Increases in the frequency of forest fires during dry periods may also be expected as a consequence of climate change induced droughts (Aragão et al., 2008). These effects enhance changes in global climate even further reported by Cox et al. (2000) and Cox et al. (2004) a strong climate feedback towards reduced vegetation and soil carbon, predicting decreases of about 128 Gt C for the period between 1860 and 2100.

Climate models predict a global temperature rise of 1 to 6 °C until the 21st century (IPCC, 2007). This warming, and the associated enhanced heterotrophic respiration, may force changes in Amazon rain forest that could lead to a future climate without analogue in the recent past (Maslin et al., 2005).

Several Global Circulation Models (GCM) simulations have predicted a die-back of rain forest in the Amazon Basin (Cox et al., 2004; Huntingford et al., 2008). The GCM showed an enhanced of soil carbon emissions, whereas reducing the size of the respiring soil carbon pool until the absolute respiration flux as a consequence of global warming. Other GCM simulations of deforestation and subsequent of part of the Amazon region indicate enhanced forest fragmentation, with associated edge effects, and an increase in temperature and decrease in precipitation (Costa and Foley, 2000; Cox et al., 2000; Oyama and Nobre, 2003; Li et al., 2006; Lewis et al., 2011). This scenario is likely to cause further mortality and ecosystem decay (Laurance et al., 2002; Hutyra et al., 2005). Recently, Nobre and Borma (2009) have identified two thresholds for the stability of the Amazon forest – climate equilibrium associated with global warming and deforestation, which are perceived as the main threats for Amazon forests. Higher risks of forest die-back or savannization over large portions of the Amazon Basin was foreseen beyond a global warming of 3–4 °C, or when deforestation in the Amazon Basin would exceed 40% of the initial forest area.

These models studies all treat the rain forest as a homogeneous cover of the dominant, well-drained upland terra firme forest, which grows on clayey to loamy soils (Oxisol and Ultisols) and covers about 70% of Amazonia (Luizão, 1996). This forest type has close to optimal photosynthesis conditions under current climate and soil moisture conditions. In central Amazonia, unconsolidated sedimentary layers have been dissected by rivers and creeks over time, which lead to formation of a landscape with rather flat plateaus being abruptly separated by steep slopes from broad swampy river valleys (Chauvel et al., 1987; Waterloo et al., 2006). Different forest types have developed within this landscape in response to topography-controlled distribution of soil types (clay content), phreatic levels and nutrient availability (Singer and Aguiar, 1986; Walker, 1987; Laurance et al., 1999). As a consequence, the Amazonian terra firme forest landscape is actually made up of several different forest types, that each may cover substantial areas. These types include seasonally inundated varzea or *igapó* valley forests along the major rivers, *campinarana* ecotone areas and riparian valley forests along the smaller rivers and low-statured *camp-*

ina forests (sensu (Anderson et al., 1975)) that occur on strongly leached sandy soils (Prance and Schubart, 1978). *Campinarana* and *campina* forests cover approximately 6% of Amazonia (McClain et al., 1997; Luizão et al., 2007). The presence of such a variation of forest types (*campinarana* and *campina*) in the landscape may reduce the overall sensitivity of its carbon cycle with respect to a faster carbon turnover or changes in ecosystem evapotranspiration and groundwater recharge (Richey et al., 2002) from all different Amazonian forests.

Drought experiments in Amazonia have been exclusively conducted in the plateau *terra firme* forests. The drought studies by Meir and Grace (2005), Nepstad et al. (2002b) and Davidson et al. (2008), suggesting that tall terra firme forest on well-drained soils are not very vulnerable to reductions in rainfall over periods of less than three years, but do respond to longer periods of drought (Lewis et al., 2011), which cause increases in tree mortality and up to 60% reduction in above-ground net primary productivity (Nepstad et al., 2007; Brando et al., 2008). (Meir et al., 2008, 2009) also reported 20–30% reduction in the leaf area index, 30–40% in transpiration rates and a 20% reduction in soil CO₂ emissions.

Whilst the well-drained Oxisols and Ultisols on slopes and plateaus have a significant clay fraction, the poorly-drained valley soils consist of predominantly of pure quartz sand (Spodosols) and have little capacity to retain water or nutrients (Singer and Aguiar, 1986; Walker, 1987). Riparian and ecotone *campinarana* forests grow on these poor sandy soils and experience almost permanently saturated soil conditions, which are maintained by a continuous flow of groundwater from slope and plateau areas, where higher groundwater levels are maintained by recharge during wet periods (Hodnett et al., 1997; Tomasella et al., 2007). Malhi and Phillips (2004) have reported that tropical forests are dynamic ecosystems of constantly shifting composition and structure. As the riparian forest, and to a somewhat lesser extend the ecotone *campinarana* forest, is adapted to near-permanent saturated soil conditions, a frequent or permanent lowering of the groundwater level due to changes in upland groundwater recharge may induce changes in the valley soil and forest composition and functioning. Indeed, Walker (1987) already perceived desertification of Spodosol areas in the Amazon as a real threat if climate should get drier.

In central Amazonia, short-statured *campina* (heath) forests are common on strongly-leached sandy soils, with a texture similar to the valley soils, but with lower phreatic levels. The main characteristics that distinguish the *campina* forest type from other forest types are its low species richness and dominance of a single or very few tree species, its unusual physiognomy – shorter stature, multi-branched and tortuous trees and the presence of bushes with scleromorphic leaves (Anderson et al., 1975; Anderson, 1981; Richards, 1996).

Cowling et al. (2003) suggest that the predominant mechanisms for maintenance of forest in cool periods are the reduced evapotranspiration and low-

ered respiration costs and that, conversely, tropical rain forests may now be near an upper temperature threshold where these physiological mechanisms become positive feedbacks that could induce forest die-back. Changes in the frequency of valley soil saturation conditions may induce changes in soil carbon and nutrient retention capacities, partly because of the poor soil nutrient retention capacity (Chauvel et al., 1987). This could potentially lead to local shifts in the composition of riparian forest towards adjacent lowland ecotone *campinarana* forest or to *campina* forest (e.g. *savannization*; (Cox et al., 2000; Oyama and Nobre, 2003) with its particular carbon cycling. If valley forest evapotranspiration is reduced due to such a shift in forest composition towards *campina* forest, this may lead to a further increase in air temperature. This is in line with reports that the tropical rain forest will increase its relative tolerance due to the temperature rise and that it is inevitable that tropical forest composition will change in response to atmospheric change, as particular species are favoured by higher CO₂ concentrations and changing climate conditions Malhi and Phillips (2004).

CO₂ gas in the soil, which eventually evades as soil respiration, can be generated by various processes. These include biotic processes, such as respiration of roots, microbes or macro-fauna in the soil. Such biotic processes are affected by temperature, as well as by moisture content. Under very dry soil conditions the ecosystem is known to have lower respiration rates (Sotta et al., 2004; Davidson et al., 2008). This mechanism was explained by Linn and Doran (1984), who observed that under very dry soil conditions the substrate diffusion through water films around soil particles to microbial active cells becomes limited. Soil CO₂ can also be generated through chemical processes, such as oxidation of soil organic matter, which may lead to either fast or slow carbon loss, depending on aeration status, soil pH and the presence of chemical components to interact with soil organic matter (and litter).

For a better understanding of the implications of climate change on the rain forest carbon cycle in central Amazonia, we need to study changes in soil carbon (Phillips et al., 2009), nutrients and heterotrophic respiration in response to lowered groundwater levels in the sandy riparian zone, which is likely to be most affected by reduced rainfall. The results presented here describe observed variations in soil respiration, groundwater levels, soil nutrients and litter decomposition in adjacent riparian forest, *campinarana* forests and in a separate *campina* forest. In addition, we report on a study of the drought response of soil in a saturated valley area where the phreatic level was artificially lowered using drainage tubes. This study allows assessment of the rate and magnitude of changes that may typically occur during extended periods of drought in sandy valley soils in central Amazonia and may provide clues on potential changes in the vegetation that is adapted to high phreatic levels as maintained by continuous groundwater flow from the plateaus and slopes.

5.2 Site description

Three measurement sites were established in the *Reserva do Cuieiras – ZF2* ($2^{\circ} 36' 32.67$ S, $60^{\circ} 12' 33.48$ W), which includes the Igarapé Asu catchment (Waterloo et al., 2006). The Reserve belongs to the *Instituto Nacional de Pesquisas da Amazônia* (INPA) and is about 70 km north of Manaus along the Manaus – Boa Vista highway. The *Reserva do Cuieiras* study site has extensively been described by Araújo et al. (2002) and Araújo et al. (2010). The topography is typical for central Amazonia consisting of rather flat plateaus (90–105 m a.s.l.) incised by broad swampy valleys (45–55 m a.s.l.) with moderately steep slopes (15–30°; (Waterloo et al., 2006)). The soils on slopes and plateaus are Oxisols and Ultisols with a significant clay fraction and cover 57% of the area (Chauvel et al., 1987; Rennó et al., 2008). The valley soils cover the remaining 43% of the area and consist of strongly leached quartz sands (Spodosols) with a low capacity to retain water or nutrients (Chauvel et al., 1987; Brinkmann, 1985; Waterloo et al., 2006). The dominant forest type on the well-drained, clayey soils on slopes and plateaus is tall *terra firme* rain forest with tree heights varying between 25 m and 45 m (Oliveira et al., 2008; Rennó et al., 2008). The poorly-drained broad valleys are covered by campinarana and riparian forests, which are characterized by tree heights of up to 25 m and have a higher abundance of palms. Phreatic levels within the research area ranged from well over 30 m below the soil surface on the plateau to less than 0.1 m depth in the riparian forest (Tomasella et al., 2007).

The drainage experiment was conducted in riparian forest using drainage and control plots established 50 m apart along the *Igarapé Asu* river in the swampy valley bottom (Fig. 5.1), where the phreatic level is usually at less than 0.1 m below the soil surface (Tomasella et al., 2007). Both plots had similar soil, vegetation composition, hydrologic and micro-climatic conditions. Ecotone campinarana forest grows on sandy soils in the zone between the riparian forest and the tall *terra firme* forest in central Amazonian catchments (Costa et al., 2005; Drucker et al., 2008). This forest type shares many plant species with riparian forest and covers an area of about 750,000 km² in the Amazon (Hutyra et al., 2005). A *campinarana* forest plot was established close to the riparian plots, about 300 m away from the *Igarapé Asu* stream. This site had a lower phreatic level at an average of 0.8 m below the surface (Fig. 5.2).

A fourth plot, representing short-statured *campina* forest growing on well-drained, strongly-leached white sands at a higher topographic position in the landscape, was established in INPA's *Reserva de Campina* ($2^{\circ} 59' 03''$ S, $60^{\circ} 03''$ W) located along the BR-174 highway 45 km north of Manaus, Amazonas State, Brazil. The size of the reserve is approximately 9 km² and the vegetation consists predominantly of lowland evergreen rain forest, but also includes patches of stunted heath forest (*campina*, 0.03 km²) and taller heath forest (*campinarana*, 0.04 km²). The *campina* forest at this site is characterized by small patches of trees and shrubs that are surrounded by areas of bare

soil *i.e.* sun *campina* sensu (Anderson et al., 1975). Tree heights range between 4 m and 7 m and between 10 m and 18 m for *campina* and campinarana *i.e.* shade *campina* sensu Anderson et al. (1975) forests, respectively (Luizão, 1996; Luizão et al., 2007). The phreatic level is deeper than at the other sites at about 3 m below the surface and the coarse grained soil is conducive to excessive drainage and is poor in nutrients (Prance and Schubart, 1978).

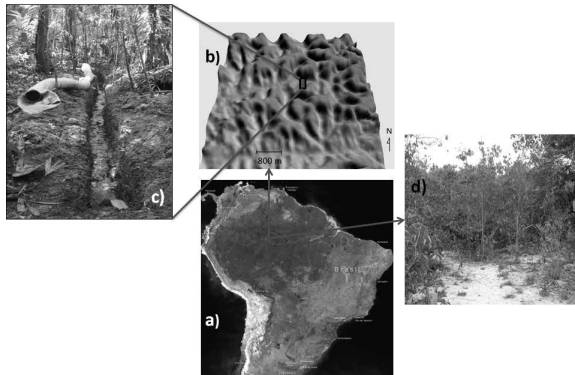


Figure 5.1: Location of research sites and trench installation: a) South America View from Google maps with arrow pointing to the ZF2 experimental site in central Amazonia, b) riparian zone (SRTM image) with broad swampy valleys and moderately steep slopes in the Cuieiras Reserve where the drainage experiment was done, c) is the drainage tube installation in trench at 1 m depth in the riparian forest plot and d) is the short-statured *campina* reserve forest.

The climate in central Amazonia does not vary much, the annual average temperature is 26.7 °C and relative humidity is about 80%. Annual rainfall, measured at the Ducke rain forest reserve near Manaus (1966–1992), amounts to 2442mm, with a standard deviation of 306mm. The rather weak dry season is between June and November (Hodnett et al., 1997; Araújo et al., 2002; Waterloo et al., 2006).

5.2.1 Measurements

Drought experiments in other parts of central Amazonia used throughfall exclusion to simulate drought *e.g.* (ESECAFLOR – Meir and Grace (2005) and Nepstad et al. (2002b)). Throughfall exclusion could not be used to simulate drought in valleys because of the constant supply of groundwater from the plateaus. Hence, we designed a method to simulate drought by artificially lowering the phreatic level without excluding rainfall. To achieve such lowering of the phreatic level, we installed two agricultural drainage tubes in parallel with 15 m spacing (Fig. 5.1). Installation of these tubes started on March 1st and finished on March 10, 2007. Each tube had a length of 30 m and 0.2 m

diameter and was installed $\bar{1}$ m below the soil surface. To avoid having to cut through large roots, the tubes were positioned at least 4 meters away from large trees. Backfilling of the trenches was done in such a way to preserve the original soil profile as much as possible.

Soil respiration was measured at all plots with a single automated soil CO₂ flux system (LI-8100, LI-COR, Nebraska, USA) with a 0.2 m diameter chamber. The system was set up to measure soil CO₂ flux, as well as soil moisture content and soil temperature (both measured at 5 cm depth) at 30-minute intervals. The instrument was consecutively used in the various plots. In each plot, measurements were made over periods of four to seven consecutive days at three locations in the plot before it was moved to the next plot. In this way information on both spatial and temporal variations could be obtained for each plot and average values reported for each plot therefore include both variations in time and space.

Decomposition of leaf litter was measured according to the litter bag method of Bockock and Gilbert (1957). We used sixty litter bags with 1 mm mesh size and larger lateral holes to allow entrance of larger insects. Each bag contained 5.75 g of dried (at 70°C) fresh leaf material collected from trees in the same plots. The experiment started at all sites on 6 December 2006, when the bags were installed on the ground surface and spread out in 12 blocks of 5 bags each. Five bags were retrieved randomly at each site every month to determine dry weight loss. A double exponential model, used earlier by Luizão and Schubart (1987) and based on that of Carpenter (1982), was used to estimate times of 50% and 95% initial mass loss rates, based on measured mass loss rates over the period of a year.

Two piezometers were installed 2 m apart in the drained plot with their screens at 0.75–1.25 m below the surface. The piezometers were installed parallel to the drainage tubes at a distance of 4 m from the tubes using hand-operated bailer boring equipment (Eijkelkamp Agrisearch Equipment, The Netherlands) on March 7, 2007. A single piezometer was installed in the control plot, with its screen at a depth of 1.62–2.12 m below the surface on March 16, 2007, about 50 m away from those in the drained plot. The fourth piezometer was installed in the *campinarana* forest at this site on the March 8, 2007.

A fifth piezometer was installed in the *campina* forest plot on January 26, 2007, with its screen at 3.62–4.12 m below the surface, reflecting the much lower phreatic level in this plot. The screen depths were such to remain below minimum dry season phreatic levels. Groundwater levels were measured weekly with an acoustic sounding device (Eijkelkamp Agrisearch Equipment, The Netherlands). In addition, groundwater samples were collected at monthly intervals from all piezometers, after flushing at least three times the tube volume using an electric pump to sample fresh groundwater.

Three sets of six vacuum-type ceramic cup soil water samplers were installed at depths of 0.1, 0.2, 0.3, 0.4, 0.5 and 0.6 m in the drained, control, *campinarana* and *campina* forest plots to obtain samples from soil moisture

and shallow groundwater. Before installation, all cup samplers were rinsed with a diluted HNO_3 solution ($\text{pH}=1$) to remove any sorbed ions and flushed with distilled water until the EC of water used for flushing fell below $2\mu\text{S cm}^{-1}$. Samples from the three sets of cup samplers in each plot were bulked, yielding single samples for each depth after each sampling event.

Water samples were analysed for Dissolved Organic Carbon (DOC), Dissolved Inorganic Carbon (DIC) and inorganic ions. DOC/DIC samples were filtered *in situ* with pre-ashed glass fiber filter ($0.7\mu\text{m}$ Whatman GFF, USA) and then stored in a 25 ml glass bottle with a Teflon lid to prevent contamination. HgCl_2 (at $300\mu\text{M}$) was added for preservation of the sample. DOC and DIC concentrations were analysed shortly after sampling on a TOC analyser (TOC5000A; Shimadzu, Japan) in the laboratory at *INPA*. Separate samples were collected for analyses of major cations (Na^+ , K^+ , Mg^{2+} , Ca^{2+} , NH_4^+) and anions (Cl^- , SO_4^{2-} , PO_4^{3-} , NO_3^- , NO_2^-). These samples were filtered using disposable cellulose-acetate filters ($0.47\mu\text{m}$, Whatman GFF, USA) and stored in high-density 1001 polyethylene bottles. Sample preservation was achieved by adding few drops of a 100 mg L^{-1} Thymol solution and then analysed on a Dionex ion chromatograph (DX500, USA).

Soil samples were collected with a hand auger set (Eijkelkamp Agrisearch Equipment, The Netherlands) at three locations within each plot to account for spatial variation. These samples were then bulked according to depth (0–0.5, 0.5–0.10, 0.20–0.30, 0.30–0.40 and 0.60–0.70 m) to limit the number of samples to be analysed. Sampling occurred both in control and plots just before installation of the drainage tubes (February 7, 2007) and sixteen months after the installation of the drainage tubes (July 18, 2008). Soil samples were collected in a similar fashion in the *campina* plot. All samples were stored in plastic bags for subsequent drying and analyses in the laboratory.

Soil sample analyses were done at the *Laboratório Temático de Solos e Plantas* – LTSP of INPA according to the methods described by Silva (1999). Analyses of %C and %N were done on finely ground soil sample triplicates using an Elemental Analyser (NA 1500 W Series 2, Fisons Instruments, UK). Extraction of exchangeable P, K, Fe, Mn and Zn was done according to the Mehlich I procedure, whereas exchangeable Ca and Mg were extracted in a 1N KCl solution as described in Silva (1999). Potential acidity (exchangeable Al+H) was measured in a 0.5N Ca-acetate solution at a pH of 7. The obtained solutions were analysed with an atomic absorption spectrometer (1100B, Perkin Elmer, USA). Soil pH and pH_{KCl} were measured in 1:2.5 soil-water and 1:2.5 soil 1N KCl solutions after equilibrating for 1 h and repeated shaking (Silva, 1999). The pH was measured with a laboratory pH meter (mPA210/mPA-210P, Tecnopeon, Brazil) calibrated with pH buffers of 4.00 and 7.00. Electrical conductivity (EC) was measured with a WTW conductivity meter (315i EC-meter, Germany), calibrated against a $1460\mu\text{S cm}^{-1}$ 0.01M KCl solution.

At the ZF2 site, 30-minute precipitation totals were measured at 51 m above ground level on top of a micro-meteorological tower with a tipping bucket rain gauge (EM ARG-100, UK; 0.2 mm resolution) connected to a

CR10X datalogger (Campbell Scientific, USA). Rainfall was measured with a similar system, but at about 3 m above the canopy in the campina forest.

5.3 Results and Discussion

5.3.1 Soil drainage status

Time series of groundwater levels in the riparian forest drainage and control plots, the campinarana plot and in the *campina* forest plot are shown in Fig. 5.2. Visual inspection of the drainage and control plots before the start of the experiment showed that the soil was completely saturated and water was on the surface at both plots. Groundwater level measurements in the control plot started only two months after installation of the drainage tubes and the phreatic level remained close to the surface at all times. The mean water level depth was 0.12 ± 0.06 m ($n=36$). Two days after the installation of the tubes (10 March 2007) the phreatic level in the drained plot had already decreased to 0.21 m and the phreatic level remained afterwards below that of the control plot, varying between depths of 0.24 m and 0.45 m below the surface, with a mean depth of 0.35 ± 0.04 m ($n=36$).

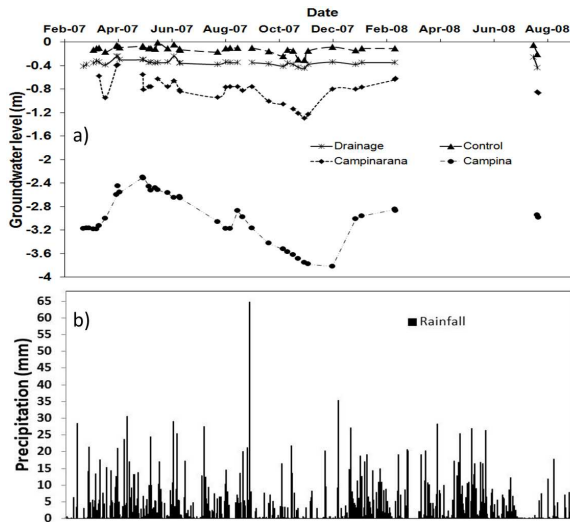


Figure 5.2: a) Changes in phreatic levels at the four research plots as observed from weekly measurements; b) hourly rainfall in the *Igarapé Asu* catchment (*ZF2*).

The installation of the drainage tubes therefore effectively lowered the phreatic level permanently by 0.23 m in the drained plot as compared to that in the control plot. Seasonality in the phreatic level is virtually absent in the

riparian forest that is fed by a constant supply of deep groundwater originating from the plateau (Hodnett et al., 1997; Tomasella et al., 2007). However, due to the sandy, permeable nature of the soil, sudden drops can occur during dry periods, such as in October – November 2007 when the phreatic level in the control plot fell to a depth of 0.31 m below the surface in response to dropping groundwater levels on the plateau. However, these dry periods generally do not last longer than a few weeks under the current climate regime in central Amazonia. The phreatic level in the drainage plot seems to exhibit less variation than that in the control plot, but also did show a response to drought in periods of low rainfall (October – November, 2007).

The campinarana plot, located between the plateau/slope and riparian area, experienced lower phreatic levels ranging between depths of 0.39 m and 1.30 m below the surface, with a mean depth of 0.81 ± 0.21 m ($n=32$). Different combinations of soil texture, drainage status and nutrient availability impact on the forest cover in central Amazonia, with tall *terra firme* rainforest on clayey soil and short statured *campina* forest on white sands occupying both ends of the spectrum in non-flooded areas with respect to the cycling of carbon, mineral nutrients and species composition (Guillaumet, 1987; Walker, 1987; Luizão et al., 2004). Our studies in riparian, ecotone *campinarana* and *campina* forests that grow on sandy soils with little capacity to retain nutrients and under varying drainage conditions, indicate that there are clear differences in litter decomposition, soil and water chemistry and soil CO₂ respiration. The phreatic level in the *campina* forest was lowest, with depths ranging between 2.3 m and 3.8 m below the surface and with a mean depth of 2.99 ± 0.42 m ($n=39$, Fig. 5.2). Phreatic levels in the *campinarana* and *campina* plots were therefore much lower than those experienced by the riparian forest and also showed a much more pronounced seasonal variation, especially in the *campina* forest plot (Fig. 5.2). The well-drained *campina* forest is an infiltration zone due to its high position in the landscape (Prance and Schubart, 1978), whereas the *campinarana* and riparian forests receive groundwater from upland areas (exfiltration zones) and therefore experience much higher phreatic levels (Hodnett et al., 1997; Waterloo et al., 2006; Tomasella et al., 2007). A slight to moderate reduction in rainfall presumably would not reduce evapotranspiration by the tall *terra firme* forest on the higher parts in the landscape, but may reduce groundwater recharge in these upland areas significantly causing a lowering of the phreatic level below the plateaus. This in turn would limit the flow of groundwater to the valley, reducing stream discharge and causing a higher temporal variation in the phreatic levels in the valley.

5.3.2 Litter decomposition rates

Litter decomposition rates observed in the riparian forest plots indicated a rapid initial loss of leaf mass in the first month (up to 25% of the initial mass loss for individual samples), followed by a more gradual decline to about 73% mass loss a year after installation (Fig. 5.3). The decomposition rate was much

lower in the *campina* forest, where only 35% of litter mass was lost in the first year. About 50% of weight loss occurred after 210 days in both control and drainage plots, whereas in the *campina* forest the 50% mass loss occurred only after 480 days. The data indicated that 95% mass loss would occur after 925 days in the control plot, 955 days in the drainage plot and only after 1825 days in the *campina* forest plot ((Fig. 5.3), it was similar to those observed by Luizão and Schubart (1987), who obtained corresponding estimates of 218 days (50% of weight loss) and 1006 days (95% of weight loss), for dry season conditions based on a 150-day measurement period.

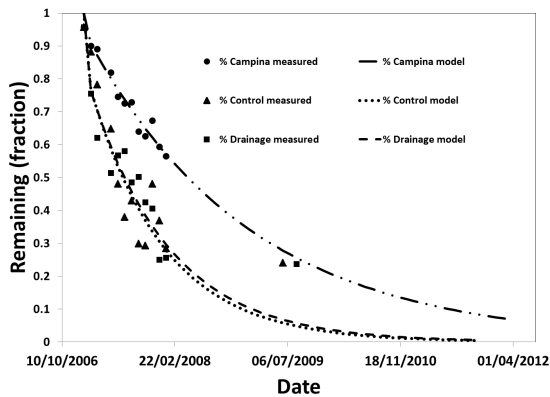


Figure 5.3: Remaining mass fraction of organic material in litterbags installed in riparian forest drainage and control plots (ZF2) and in the *campina* forest plot.

Luizão and Schubart (1987) observed that litter decomposition was faster on the plateau than in the valley of this *terra firme* landscape and that decomposition rates were higher during the wet season (50% mass loss in 32 days) than during the dry season. The latter observation is not supported by our measurements that started early in the wet season and continued throughout the dry season. According to Luizão et al. (2004), litter turnover rates on plateau and slope were not significantly different from those in the valley in spite of lower N concentrations in the upper soil layer in the valley, which they attributed to other processes, such as leaching, removing litter from the soil surface.

A permanent decrease in phreatic levels or a change in the frequency and lengths of droughts may impact the local decomposer community, and thereby decomposition rates of leaf litter. However, no significant differences were found between decomposition rates in the drained plot, where the phreatic level was artificially lowered by about 0.3 m, and the control plot (Fig. 5.3). This may have several causes. First, the decomposition experiment started four months before drainage tubes were installed and about 40% of the initial litter mass had therefore already been lost before the phreatic level was low-

ered. Furthermore, the size of the drained plot may have been too small (edge effects, micro-climate and vegetation) and the period of study after drainage started (1 year) too short for adaptation of the decomposer community to the new drainage conditions in the plot.

There may be several reasons for the slow decomposition observed in campina forest. Tree species in campina forest do have sclerophyllous leaves that may be more difficult to decompose and studies have shown that the decomposition activity of fungi is also suppressed in this environment, where ectomycorrhizae are instrumental in the cycling of nutrients (Singer and Aguiar, 1986). Walker (1987), Luizão and Schubart (1987) and Luizão et al. (2004) have reported that even the high annual rainfall, which percolates rapidly through the nutrient-poor sandy soils, may not be sufficient to transport sufficient nutrients for the synthesis of adequate quantities of structural proteins (endoplasmatic reticulum, membranes, histones, etc.) in order to balance intense assimilation. This could be a cause for a relative excess of cellulose and lignin in the leaves. The first step in litter decomposition is leaching, when nutrients and organic compounds dissolve in water and move into the soil. Compounds that are easily leached from litter include potassium, sugars and amino acids. Subsequent leaf litter weight loss is related to termite activity in *terra firme* rain forest (Luizão and Schubart, 1987), which can also degrade such resistant substances as lignin (Butler and Buckerfield, 1979). When litter becomes increasingly fragmented over time, it reveals new surfaces for bacteria and fungi to attack (Chapin et al., 2002). The presence and activity of termites may be low in nutrient-poor campina forest as compared to those in other rain forest types, which may slow down litter fragmentation and decomposition in campina forest, causing low carbon levels in the soil and high DOC concentrations in soil moisture and groundwater, as was observed in this and other studies in central Amazon (McClain and Richey, 1996; McClain et al., 1997). Litter fragment decomposition contributes significantly to CO₂ production in the soil, accounting for about 27% of soil emission (Wang et al., 1999), which is low at this site. Su (2005) reported that litter decomposition rates decrease over time due to labile compounds leaving the litter being attacked quickly. This process leaves a residue of less readily decomposable substances (Su, 2005), with potentially high C:N ratios (McClain et al., 1997). High nitrogen content in litter is considered a factor affecting the speed of decomposition by micro-organisms. Pate and Lyzell (1990) shows that cellulose degradation is also a nitrogen-limited process and will increase with the nitrogen content of litter. Vitousek and Sanford Jr. (1986) compared foliar and fine litterfall nutrients of various rainforests and showed that nitrogen and phosphorus appear to cycle less in campina forest than in other lowland forests.

5.3.3 Groundwater and soil chemistry

Average of chemical analyses of the groundwater obtained from piezometers in control and drainage plots are given in Table 5.1, whereas time series of pH

and DOC concentrations in groundwater for all plots are shown also in Fig. 5.4. DOC levels were consistently highest in groundwater below the campina forest, at 20-25 mg L⁻¹, and lowest in the campinarana plot, varying seasonally between 2 and 15 mg L⁻¹. The pH was invariably low at piezometer measurements in situ, ranging between 3.5 and 4.7, with a single minimum of 3.21 being observed in the drainage plot at the start of the drainage experiment in both piezometers. In both drainage and control plots the *EC* was positively correlated with the DOC concentration being $R^2=0.90$ and 0.84 , respectively, whereas the pH correlated negatively with DOC concentration $R^2=0.26$ and 0.60 , respectively (Fig. 5.4). Concentrations of K⁺ and NO₃⁻ remained invariably low.

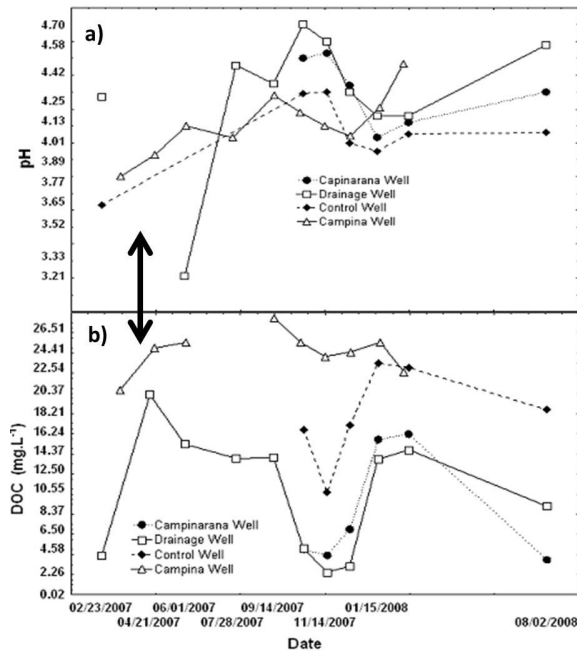


Figure 5.4: Temporal variation of in pH (a) and DOC (b) concentrations in groundwater (wells samples) in all four research plots. The dual arrow indicates the start period of drainage experiment.

The temporal variation in DOC levels observed in the campina plot is slighter than those in the control, drainage and campinarana plots. The fluctuations in the DOC concentrations of valley groundwater are presumably caused by seasonal variations in the rate of up-welling of deeper (plateau) groundwater, in which DOC concentrations are invariably low (<5 mg L⁻¹) according to (Waterloo et al., 2006). Groundwater in the control plot had the highest DOC concentrations, whereas below the drainage and campinarana plots were con-

Table 5.1: Chemical analysis of groundwater in the drainage and control plots.

Control piezometer, 1.62–2.12 m depth							Drainage piezometer, 0.75–1.25 m depth					
Period	pH	DOC (mg L ⁻¹)	DIC (mg L ⁻¹)	NO ₃ (mg L ⁻¹)	K (mg L ⁻¹)	EC (μ S cm ⁻¹)	pH	DOC (mg L ⁻¹)	DIC (mg L ⁻¹)	NO ₃ (mg L ⁻¹)	K (mg L ⁻¹)	EC (μ S cm ⁻¹)
23/02/2007	3.63	-	-	-	0.00	24.2	4.27	3.90	2.25	0.01	0.03	17.0
After drainage installation (31 March 2007)												
21/04/2007	-	-	-	-	-	-	-	19.87	0.76	0.01	0.35	-
01/06/2007	-	-	-	-	-	-	4.21	14.97	4.32	0.01	0.34	25.1
01/08/2007	-	-	-	-	-	-	4.46	13.54	3.68	0.06	0.22	21.0
14/09/2007	-	-	-	-	-	-	4.35	13.64	0.75	0.05	0.40	23.2
19/10/2007	4.29	16.38	8.96	-	0.23	32.9	4.70	4.63	9.17	0.08	0.61	13.5
16/11/2007	4.30	10.14	3.77	0.03	0.26	18.4	4.60	2.26	3.86	0.06	0.41	13.7
13/12/2007	4.00	16.85	7.56	0.03	0.20	29.2	4.30	2.85	8.44	0.04	0.3	13.5
15/01/2008	3.95	23.03	7.71	0.02	0.21	35.8	4.16	13.43	5.38	0.13	0.25	22.4
21/02/2008	4.05	22.56	6.73	-	0.09	35.5	4.16	14.4	4.24	0.04	0.2	25.2
02/08/2008	4.06	18.40	6.42	0.01	0.97	34.0	4.58	8.79	6.07	0.17	0.46	21.0

sistently lower by about 7 mg L^{-1} . The *EC* in groundwater varied between 13 and $36 \mu \text{ S cm}^{-1}$ in the control and drainage plots. These differences could not be attributed to the impact of drainage as the initial pH and *EC* measurements suggest that DOC concentrations were already higher in the control plot than in the drainage plot at the start of the drainage experiment (Fig. 5.4).

The drainage study did not reveal significant changes in groundwater and soil chemistry conditions, excepting for the marked changes in *C* and *N* contents of the top 5 cm of the soil in the drainage plot as compared to those in the control plot. The soil acidity, defined as H^+Al , is high in the riparian and campina forests. The low pH values observed in soil moisture in the drainage and control plots did not seem to be associated with corresponding increases in DOC concentrations and must therefore be caused by different soil processes, such as the removal of H^+ ions from exchange sites by other cations. Luizão et al. (2007) did observe that H^+ was the dominant ion, rather than Al^{3+} , on the exchange complexes in campina forest and suggested that H^+ ion toxicity is a major growth-limiting factor for non-adapted plants in heath forest soils in central Amazonia. Our study indicates that the riparian forest soils may have this low Al (and Fe) sesquioxides content in common with the campina soils, which could have implications for the forest succession in a drier climate. The yellowish/orange color of the sand below campinarana forest suggests that these soils do have a higher Al (and Fe) sesquioxides content.

The shallow groundwater, sampled with the ceramic cup samplers, showed less seasonal variation in DOC concentrations and pH than that observed in the deeper piezometer samples (Fig. 5.4). Water extracted from the 0-10 cm soil depth in the drained plot had a comparable DOC concentration as those measured deeper in the soil in the control plot. However, in the drained plot DOC concentrations decreased in the subsoil (below 10 cm) and were consistently lower than those measured at the same depths in the control plot (see Fig. 5.4). Topsoil DOC concentrations also seem to respond to heavy rainfall in both plots. As this was the case from the start of the experiment onwards, it cannot be attributed to the artificial drainage.

The forest plots in exfiltration zones showed higher groundwater levels, higher soil nutrient contents and soil respiration rates, and faster litter decomposition rates. In spite of this, DOC levels in the shallow groundwater and stream water at these sites are also high and may reach levels locally that are similar to those presently observed in the campina forest groundwater. This suggests that the litter decomposition process that provides DOC in the valley is different from that acting in terra firme forests on slopes and plateaus in the central Amazonian landscape, where DOC levels in groundwater remain low Luizão et al. (2004), Waterloo et al. (2006) and Marques (2009) showed that soil *C* and *N* contents were significantly higher in the clayey topsoil (0–10 cm) on plateau and slope, than in the sandy valley, which they attributed to higher biological activity and higher intensity of mineralization processes at the former sites. Ion concentrations are invariably low in soil moisture and groundwater, as is the pH which varies between 3 and 5.

In spite of the lower DOC concentrations in soil moisture in the drainage plot for most of depths (Fig. 5.5), the pH was comparable to that in the control plot, but showed a somewhat lower seasonal variation never reaching the extremely low values observed in the control plot (pH of about 2.8 in June 2007, Fig. 5.5).

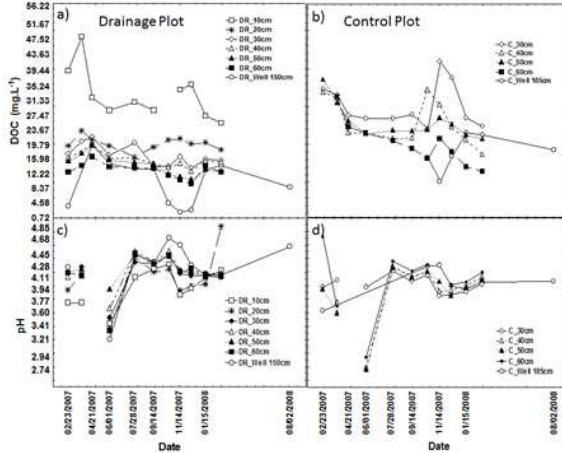


Figure 5.5: Temporal variation in pH and DOC concentrations from cup sampler (shallow) and piezometer (deep) groundwater in drainage (a and c) and control (b and d) plots.

A comparison of the DOC concentrations measured at all sites is given in Fig. 5.6. We have tested (F -test) differences between samples collected at all sites. The DOC concentrations were lowest in the riparian forest plots, whereas the highest values were observed in the *campinarana* forest. However, there was a significant difference between the DOC concentrations in soil moisture at 0.3 m depth in control and drainage plots ($n=16$, $F=13$, $p=0.00002$). The DOC concentration at the surface 0–10 cm layer in the drainage plot was similar to that measured at 0.3 m in the control plot. DOC concentrations in the control (at 0.3 m depth; $n=15$, $F=4.27$, $p=0.05$) and drainage (at 0.2–0.3 m depth; $n=25$, $F=3.8$, $p=0.04$) plots were significantly different from those (10–40 cm) in the *campinarana* plot. DOC concentrations in the *campinarana* plot showed a large variation and resembled those measured in the *campina* plot more than those in the riparian forest plots. It should be noted that the coarse sandy *campina* forest soil often too little moisture for soil moisture extraction with the ceramic cup samplers. Samples could therefore only be collected shortly after high rainfall events.

Table 5.2 shows the chemical properties of the soils in the drainage, control and *campina* plots. The soil samples collected in the control and drainage plots at the start of the study were inadvertently lost and the results shown in

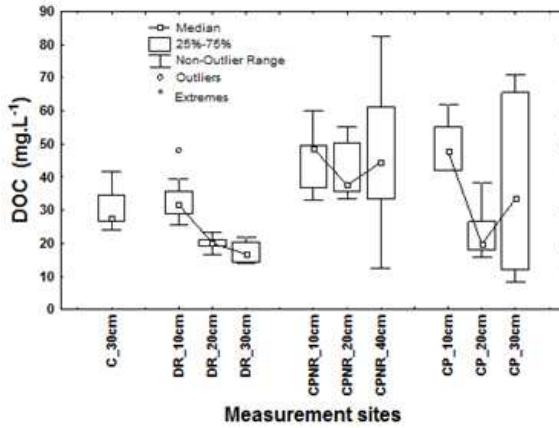


Figure 5.6: Variation in DOC concentrations in shallow groundwater sampled with the ceramic cup samples at depths between 10 to 30 cm below the soil surface in control (*C*), drainage (*DR*), *campinarana* (*CPNR*) and *campina* (*CP*) plots.

Table 5.2 are for soil samples collected in July 2008, at the end of the drainage experiment. All samples have low exchangeable cation concentrations, consistent with the sandy substrate and high leaching environment of these valley rain forests, and showed a decline with depth. The poor retention capacity of the valley soil can be due to the virtual absence of clay minerals (<2%) and the low organic matter content (Table 5.2). As expected %C, %N and exchangeable cation concentrations in control and drainage plots are rather similar. The C:N ratios remained fairly constant with depth down to 0.7 m, averaging to 24 ± 2 over the profile in the drained plot and to 28 ± 3 in the control plot. The exception is the top 0-5 cm soil layer in the drainage plot, which had extremely low %C (0.07%) and %N (0.00%) contents as compared to those observed in the same layer in the control plot (1.38% C and 0.05% N, Table 5.2). The change in carbon and nitrogen contents were in line with visual observations of a change in top soil colour in drainage plot (from dark grey to white) during the study, whereas no change in soil colour was observed in the control plot. Carbon and nitrogen difference in the first 0.05 m comparing drainage to the control plot after 16 months of treatment was about 10020 kg C ha and 365 kg N ha, respectively, with a measured bulk density of the top soil of 1400 kg m^{-3} . The analyses suggest that drainage did cause a marked decrease in the organic matter content of the top 5 cm of the soil, but not in deeper layers. Exchangeable macro-nutrient cations (Ca, Mg, P and K) remained at similar levels in drainage and control plots throughout the soil profiles (Table 5.2). The dominant ions on the exchange complexes were, in descending order, $\text{Al}^+\text{H} > \text{K} > \text{Ca} > \text{Mg} > \text{Fe} > \text{P}$.

Table 5.2: Soil nitrogen, carbon, exchangeable cations and pH at the different research sites sampled in February 2007 (CP plot) and July 2008 (DR, C plots).

Samples	N	C	Ca	Mg	Al ⁺ H	K	P	Fe	Zn	Mn	pH _{KCL}	pH _{H2O}
	%	%	mg kg ⁻¹	mg kg ⁻¹	cmole kg ⁻¹	mg kg ⁻¹	mg kg ⁻¹	mg kg ⁻¹	mg kg ⁻¹	mg kg ⁻¹		
DR - 0-5 cm	0.00	0.07	26.50	5.80	0.23	71.00	5.98	11.9	1.50	7.30	-	4.31
DR - 5-10 cm	0.15	3.94	14.50	1.80	0.16	25.80	2.02	9.70	0.50	2.50	-	4.38
DR - 10-20 cm	0.11	2.48	13.50	11.00	0.18	22.10	1.49	12.5	0.50	1.40	-	4.42
DR - 20-30 cm	0.11	2.39	9.00	7.50	0.14	12.70	1.22	10.6	0.40	0.80	-	4.58
DR - 30-40 cm	0.07	1.81	7.00	5.90	0.12	8.80	1.29	7.50	0.20	0.40	-	4.59
DR - 60-70 cm	0.05	1.25	6.50	4.60	0.12	9.50	0.40	10.8	0.30	0.40	-	4.50
C - 0-5 cm	0.05	1.38	19.50	5.10	0.18	46.30	2.25	13.0	0.80	1.40	2.69	4.25
C - 5-10 cm	0.13	3.40	38.00	8.90	0.22	53.80	4.43	16.6	1.20	1.90	2.68	4.29
C - 10-20 cm	0.23	5.94	13.50	1.90	0.19	28.90	2.38	13.6	0.70	1.10	2.83	4.36
C - 20-30 cm	0.11	2.80	14.50	1.30	0.19	21.20	1.65	11.8	0.60	0.90	2.95	4.55
C - 30-40 cm	0.09	2.59	10.00	8.40	0.17	26.60	0.89	12.4	0.50	0.40	2.84	4.55
C - 60-70 cm	0.07	2.22	9.50	6.90	0.11	11.60	0.59	11.8	0.40	0.40	2.83	4.60
CP-0-5 cm	-	-	28.2	2.10	0.55	1.77	1.19	9.75	0.23	0.83	2.83	4.27
CP-5-10 cm	0.04*	0.20*	1.00	0.00	0.50	1.65	0.90	11.5	0.15	0.68	3.20	4.67
CP-10-20 cm	0.037*	0.28*	0.50	0.00	0.15	1.67	0.37	11.8	0.23	0.25	3.46	5.05
CP-20-30 cm	0.03*	0.15*	0.80	0.00	0.10	0.55	0.35	11.1	0.20	0.28	3.93	5.31
CP-60-70 cm	0.017*	0.08*	0.60	0.00	0.05	0.37	0.28	9.38	0.10	0.30	4.23	5.37
CP-90-100 cm	0.01*	0.07*	0.70	0.00	0.05	0.80	0.27	9.25	0.15	0.30	4.20	5.28

DR: Drainage (July 2008)

C: Control (July 2008)

CP: Campina forest

* Campina data from Luizão (1996)

Exchangeable cation concentrations were lowest in the sandy campina soil, which also had the highest soil pH and pHKCl values. Al⁺H were again dominant on the exchange complex, but this was followed by Fe and very low levels of, in descending order, Ca > K > P > Mg. Observations of %C and %N made by Luizão (1996) in campina forest soil indicated that %N was similar to that measured in the riparian forest plots, whereas %C was much lower and close to that found in the top soil of the drained plot (Table 5.2). The soil C:N ratio also remained fairly constant throughout the profile in the campina forest, but was significantly lower at 9.4 ± 1.0 (Luizão, 1996).

5.3.4 Soil CO₂ respiration

Soil CO₂ respiration rates were significantly different between forest types and drainage and control plots (Fig. 5.7). The *campina* forest exhibited very low soil respiration rates at $1.2 \pm 0.3 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$, whereas *campinarana* forest showed the highest rate at $5.7 \pm 1.8 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$. Both forest types grow on soil with similar coarse sandy texture and with phreatic levels generally below 0.8 m. These forests do differ in the presence (*campinarana*) or absence (*campina*) of a root mat and continuous litter layer, which may explain the large difference in autotrophic and heterotrophic soil respiration rates. The riparian forest on near-saturated sandy soil (before drainage tube installation) respired at an intermediate rate of $3.6 \pm 0.6 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$, which is similar to that measured in the tall *terra firme* forest on the plateau.

At the start of the drainage experiment soil respiration was measured at two different locations within the drained area between 18 November and 5 December 2006, followed by measurements at a single location in the control plot between 5 and 12 December, 2006, and at a second location between 7 and 9 February, 2007. Measured soil respiration rates were very similar in control and drainage plots at 3.6 ± 0.6 and $3.7 \pm 0.7 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$, respectively (Fig. 5.7). A comparison of the soil respiration rates in the drainage and control plots indicated significant reductions of 42% ($2.1 \pm 0.2 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) in the former with respect to the initial value after 8 months of drainage and 76% ($0.8 \pm 0.2 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) after 16 months of drainage, whereas soil respiration in the control plot remained at or above the initial value at $5.1 \pm 0.8 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ after 8 months and $3.7 \pm 2.8 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ after 16 months (Fig. 5.6). Another drought experiment also found 20% of soil CO₂ emission decrease due to soil moisture decrease (Meir et al., 2008, 2009). This suggests that the drainage of the riparian forest soil caused a significant decrease in soil CO₂ respiration to the values observed in the campina forest plot.

This study is the first to provide information on soil respiration in campina forest and on the impact of drainage of sandy valley soil on soil respiration rates. The drainage experiment revealed that the largest changes are likely to occur in the top soil where the carbon and nitrogen contents decreased markedly after artificial lowering of the phreatic level. The reduction in the

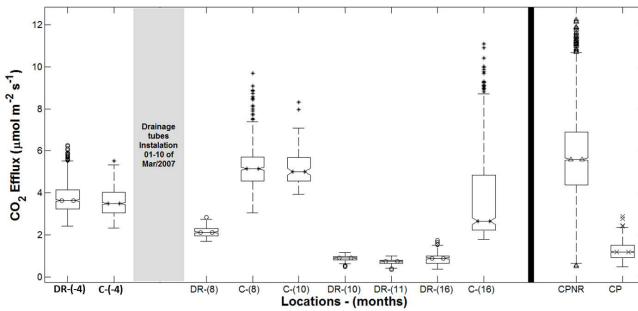


Figure 5.7: Boxplot of soil CO_2 respiration measurements at the different forest plots: Drainage plot (*DR*), control plot (*C*), *campinarana* plot (*CPNR*) and *campina* plot (*CP*). The grey rectangle separates measurements done before and after drainage tube installation and the numbers between brackets indicate the month during which measurements were made in 2007, where month 4 represents the start of the drainage experiment. The black vertical bar separates *campinarana* and *campina* measurements from those made in control and drainage plots. Fluxes in both *campinarana* and *campina* forests were measured over periods of several weeks in the wet, as well as in the dry season.

carbon and nitrogen content in the top soil were matched by a strong reduction in the soil respiration rate as compared to that measured in the same plot before drainage and in the control plot. The observations suggest that changes in topsoil carbon and nitrogen content occur within a few months after a lowering of the phreatic level and that this has a pronounced effect on the soil respiration rate. Such fast response may be caused by a combination of the rapid flushing of fine particulate and colloidal carbon in the soil pores to lower soil layers with infiltrating rainfall (Marques, 2009) and removal of particulate carbon in the form of CO_2 by enhanced decomposition rates in the better aerated top soil after drainage. Impacts of drainage on other nutrients and on those in deeper soil layers were not observed (Table 5.2, Fig. 5.5 and Fig. 5.6) and may only become apparent after a more extended period of drainage.

The soil respiration rate in the drained plot became as low as that in the *campina* plot, which presumably reflects the low *C* and *N* contents in the topsoil after drainage, where most of the soil CO_2 respiration is produced (Hanson et al., 2000). These soil carbon and nitrogen levels were much lower than those in the control plot where soil CO_2 respiration remained constant and at a rate similar to that in the drainage plot before the start of the experiment. Lambers et al. (1998) and Xu et al. (2004) have reported that nitrogen and carbon availability in the soil are important for both plant growth and the production of CO_2 . Hence, the reduction of soil CO_2 respiration in the drainage plot may be an indication that large-scale drainage of these sandy valley areas,

and associated decrease in the top soil carbon and nitrogen contents, may lead to a change in the carbon cycle towards that of campina forest. A lowering of the phreatic level in this zone and corresponding infiltration of rain water could affect soil respiration fluxes rapidly, as was observed in the present study (Fig. 5.7).

Although a modest decrease in the phreatic level may not cause hydrological stress on the vegetation, the almost daily precipitation events (Fig. 5.2b) may eventually cause a further flushing of particulate organic carbon and nutrients from the topsoil to deeper layers in the valley. However, with the low carbon contents in our sandy soils, this is less likely to be very important here. Finally, because the phreatic level is at or very close to the soil surface in riparian forest, degassing of groundwater with high $p\text{CO}_2$ might also contribute to measured soil respiration. Further research is required to discover if degassing of shallow groundwater is an important factor in this area or if biotic processes are dominant. Measuring such individual contributions of biotically, chemically and physically produced CO_2 to the soil respiration flux is extremely difficult, but would provide valuable insight into the importance of the various processes generating CO_2 and would also be important for predicting changes in soil respiration as a consequence of a drier climate.

5.4 Conclusions

Forests on sandy soils cover a significant part of the central Amazonian landscape. The present study shows that depending on the drainage status of sandy soils, a large natural variation exists in soil carbon, nitrogen and exchangeable cation concentrations, soil water chemistry and litter decomposition rates. Litter decomposition and soil CO_2 respiration rates are very low in campina forests, where the capacity of the coarse sandy soil to retain nutrients is extremely low. The highest soil CO_2 respiration rates within the terra firme landscape are found in ecotone campinarana forests that form the transition from riparian forests on nearsaturated sandy soils to tall terra firme forests on slopes and plateaus with clayey soils.

Under present climate conditions the vegetation in the riparian forest is not likely to experience significant periods of droughts. Artificial drainage of the riparian forest plot, lowering the phreatic level by a modest 0.3 m, did not cause apparent changes in the soil nutrient content, chemical composition of soil water or litter decomposition rates within the study period of one year. However, marked reductions were observed in the top soil carbon and nitrogen content, as well as in the soil CO_2 respiration rate, which decreased to the levels observed in campina forest. The drainage study therefore indicates that the riparian forest soil is very sensitive to changes in the phreatic level. It remains unclear if slower changes in soil and water chemistry in the subsoil will become apparent within the next few years. This suggests that the characteristics of drained riparian forest soil show a transition towards that of campina for-

est, with its low nutrient content and very low soil respiration rate. However, shifts in vegetation species and a future development of a root mat, such as exists in the campinarana forest, may result in increased soil respiration rates. It is as yet uncertain what the direction of succession of drying riparian forest will be and continued monitoring of the drained and control plots may provide valuable clues in this sense.

The large difference in soil respiration rates between campina and campinarana forests imply that knowledge about the likelihood of riparian forest to be succeeded by one type or the other is crucial for determining the future carbon budget of terra firme rainforest ecosystem experiencing a drier climate.

Chapter 6

Estimate of the Leaf Area Index (LAI) and Biomass in pasture in the state of Rondônia – Brazil⁵

6.1 Introduction

Over the last decades, a vast area of the Amazon rain forest has been intensively deforested for agriculture, for agro-forestry systems and mainly for cattle raising activities. The transformation of these areas to pasture usually starts with logging of the primary forest, followed by burning of the slash and the introduction of crops for harvesting with associated degradation of the soil (Fearnside, 1989). FAO (2005) has reported about the increasing importance of tropical pastures in the management of the water resources and local models of vegetation interaction with the atmosphere and its relation to climate change. However, information about the spatial variation or long-term variations in biomass, leaf area index (LAI, m^2 of one-sided leaf area per m^{-2} soil surface) as well as the specific leaf area (SLA) values, and grass height is still very scarce. About impacts of human activities, such as fire management, or through studies on the spread of weeds and pests, some facts are known about the foliage coverage in pasture. Under natural conditions, the Leaf Area Index (LAI), the height and biomass of certain grass species vary with seasonal and annual variations in climate leading to changes in the soil moisture

⁵The contents of this chapter have been published in Portuguese as F. B. Zanchi, M. J. Waterloo, L. J. G. Aguiar, C. von Randow, B. Kruijt, F. L. Cardoso, A O. Manzi. Estimativa do Índice de Área Foliar (IAF) e Biomassa em pastagem no estado de Rondônia, Brasil. *Acta Amazonica*. vol. 39(2): 335–348, 2009. (doi:10.1590/S0044-59672009000200012)

status. During extended periods of drought the plants respond to water stress and limitations to transpiration through the closure of stomata, followed by die-back and loss of leaves, and in extreme cases, by the complete senescence of above-ground plant parts (Waterloo, 1994). For pastures under natural conditions, the plant die-back can be part of an annual process provoked by the flowering and production of seeds at the end of the rainy season. This can lead to considerable reductions in the biomass, in grass height and LAI, as reported by Waterloo (1994).

Parameters of vegetation are usually obtained through short sampling campaigns, capturing only the seasonal variations of dry and wet periods (Roberts et al., 1996). The use of the parameters of the vegetation height and LAI are often convenient in climatic and hydrologic models for estimating energy flows (Ashby, 1999). These parameters modulate and regulate energy, as well as gas, exchanges (e.g. CO₂ and H₂O) between vegetation and atmosphere through their impact on aerodynamic and surface conductances. Although modelling the variation in these plant parameters may provide a good indication of the seasonal variation for natural vegetation, it is less well for describing impacts of cattle grazing, fire or other disturbances in the pasture, which usually take place on short time scales (days to several weeks). This study, which is based on monthly destructive sampling measurements, aims to provide better insight into the variation in the pasture ecosystem land surface cover.

Under natural conditions, with or without the influence of cattle grazing on the pasture, changes in the seasonal variation in pasture height, biomass and LAI can be very significant. Waterloo (1994) reports that in the Fiji Islands the seasonal LAI change ranged from 0.2 at the end of the dry season to 2.1 in the wet season, whereas the height ranged from 0.9 to 1.79 for non-grazed *Penisetum polystachyon* grassland. As for the Amazon region, these variations may be less due to the shorter dry season and the deeper soils, which generate less water stress in the plants because of the higher availability of soil moisture throughout the year. However, there may be large reductions in height and LAI during the rainy season if cattle grazing occurs in the area, which can delay and enhance growth to cause an opposite effect during the dry season, allowing the pasture to remain green even during the dry season.

Several attempts have been made to model the energy balance in pastures of the Amazon region. Based on the data of the ABRACOS Project (Nobre et al., 1996), Wright et al. (1996) recommended the use of simple modelling of LAI variation of the pasture using an annual cycle to model the growth of plants, their senescence and corresponding change in albedo with the dry season. Viterbo and Beljaars (1995); Garratt (1992) emphasized that the energy balance, as well as the latent and sensible heat fluxes, influence the simulations by atmospheric models, given that the modelling studies depend on the accuracy of the representation of processes on the surface. These processes are directly linked to the functioning of the ecosystem (e.g. pasture, forest).

Modelling of the land cover can provide an indication of the temporal and spatial variation, but there is yet not much field validation data to allow for

incorporation of rapid vegetation height changes, changes in biomass and LAI, because of changes in pasture management including periods with grazing, followed by regrowth.

The present study is part of the Large Scale Experiment in the Amazon Biosphere and Atmosphere – LBA (Nobre et al., 1996; Andreae et al., 2002). The monthly measurements were based on the measurement methods used in the two-year period of data collection in the ABRACOS Project (Gash et al., 1996). The gathered data adds towards the climate, energy and the CO₂ exchange data, which were collected during the rainy season of Atmospheric Mesoscale Campaign (AMC) in 1999, for assessing regional climate change and for the modelling of carbon sequestration within the LBA program (Andreae et al., 2002).

One of the objectives of the LBA experiment is to promote obtaining long time series of scientific data, this to increase the capacity of regional climatic modelling in view of land use, land cover and soil changes in the Amazon region. The present study contributes towards this objective by providing the first continuous six-year time series of pasture vegetation height, biomass and LAI, from 1999 to the beginning of 2005, for a pasture at near Ouro Preto d'Oeste, and wet season data for a less intensely grazed pasture near the city of Rolim de Moura, Rondônia (Brazil).

6.2 Site description

The study area is Fazenda Nossa Senhora (FNS; 10° 45' 44" S, 62° 21' 27" W), which is located near the city of Ouro Preto d'Oeste, some 50 km Northwest of the city of Ji-Paraná in the State of Rondônia (Brazil). The research field is part of a large flat area of pasture in the region, covering an area of approximately 4 km² at an altitude of 220 m above sea level (Culf et al., 1996). The soil depth can reach down to 7 m and the reddish brown soil has a fine-grained texture at the surface, with an increasing clay fraction with depth. The surface has been compacted by cattle trampling, as indicated by the soil density, which increases from 1.50 1.55 g cm⁻³ near the surface to 1.65 g cm⁻³ at 40 cm depth, decreasing with depth to 1.46 g cm⁻³ at 100 cm. The soil is a weathering product from the bedrock and has fragments of quartz and stones. On the surface, the soil has a dark-brown color, changing to reddish-brown with depth. Presence of manganese was observed below 2 to 3 m, which is related to the seasonal variation of the phreatic level. Houghton et al. (1987) classified the soil as a red-yellow podzol.

The forest in this area was logged and burned in 1977 to open fields for rice, bean and manioc crop cultivation. When the production in the area decreased due to soil degradation, *Urochloa brizantha* pasture (also known as *Brachiaria brizantha*) (Nobre et al., 1996) was established in the beginning of the 1980s. The pasture had been burned four times, with the burn in 1994. The pasture is composed by 99% of *Urochloa brizantha* and has only minor

cover of other common grass species in the region, *i.e.* *Panicum maximum* and *Urochloa humidicula*.

The roots of *Urochloa brizantha* were observed to reach a depth of 3 m in an auger hole. The cattle rotates through the pasture fields every 3 to 4 months, based on a local pasture rotation management system, which would not degrade the pasture (feeding an average of 2 to 3 animals per hectare), and guarantees to keep enough vegetation for dry season grazing. With this management method, the grass height reaches only 0.5 m height, which is below its potential height.

The annual average temperature varies between 23 to 24° C, with lower temperature events (passage of cold fronts) occurring in June and July. The monthly average precipitation is about 200 mm from November to April and drops down to generally less than 20 mm per month between June and August, being less than 5 mm in July (Nobre et al., 1996). Oliveira et al. (2001) reported that the annual precipitation for the region reached 2295 mm in 2000. The relative humidity is of 80% during the rainy season and drops to 70% in the dry season. The phreatic level varied between 3 m and 4 m below the surface in the rainy period and dropped to 6.5 m depth during the dry season. The winds are predominantly from the North, but come from the South in June/July, which is related to the low temperature events (Zanchi et al., 2002).

The research site in Rolim de Moura (RDM) is located 5 km from the broad valley of Pedrinha River (11°42' S, 61°46' W, 224 above sea level). This site belonged to the local government and the area had been invaded by landless families that used it for agriculture activities. There was less intense grazing in this pasture in comparison to that at FNS. The pasture is characterized by a mix of *Urochloa brizantha*, *Urochloa humidicula*, other types of grass, bushes and a few palm trees in an area of 240 ha. The soil is less deep (2 m) and has a finer texture in comparison to that at FNS. The climate in RDM is very similar to that of Ouro Preto d'Oeste. The RDM samples were collected exclusively in the rainy season of 1999 during the AMC campaign.

6.3 Material and methods

LAI measurements can be made through direct and indirect methods. The indirect methods are characterized by measurements of total radiation that penetrate the interior of the foliar coverage. One of the ways for capturing the radiation fraction is through a digital camera attached to a fish-eye lens, as described by Norman and G.S. (1989). In this study, the direct, destructive method was used for biomass and LAI measurement in the pasture of FNS, according to the methods described by Roberts et al. (1996) and Waterloo (1994). This is a time consuming method, but also one of the most reliable methods because it uses direct field information on the quantity of the biomass, and therefore does not depend on radiation scattering theory and the LAI calculation does not have to be corrected for inclusion of stems in the

measurements.

The LAI, SLA, quantity of living biomass and dead material were collected in five randomly chosen 1-m² plots and sampling was done once a month. Between February 1999 and January 2005, a total of 300 1-m² samples were collected at FNS. The percentage of foliage coverage, pasture height and ground water level were also measured at the same time as sampling occurred.

The sampling started with collecting of the dead material (litter) from each of the five the marked 1 m² areas. The removed material was stored in plastic bags, with good sealing, avoiding loss of moisture. The remaining living material was cut off and, whenever other species but *Urochloa brizantha* were found, they were stored in different bags to separate species from the other. All the material was pre-weighed (fresh weight) at the Universidade Federal de Rondônia (UNIR, Brazil) in the city of Ji-Paraná right after the sampling and then placed in an oven at 80°C for a minimum of 72 hours. The dry weights were then measured to obtain biomass and moisture of the material.

To obtain fractions of the dead and living material, as well as the stem to leaf fractions, 20 plants in the vicinity of the sampled area were randomly sampled. The plants were cut at soil level, obtaining all its parts (green leaf, dead material and stems). This was done at all 5 sample sites. The 100 plants were stored separately in plastic bags and the living (green) leaves, dead parts and stems were separated and sorted in the laboratory. The stems and the dead material were dried (80 °C) and weighed, whereas the contours of the green leaves were first traced on paper for calculation of the area. Hence the leaf area for each group of twenty plants was obtained. The leaves were then dried and weighed for determination of the SLA.

To calculate the LAI the fraction of fresh leaves on the plants was used and combined with live biomass and specific leaf area according to the following equation:

$$LAI = \frac{M_{leaf}}{(M_{leaf} + M_{stem} + M_{dm})} \cdot M_{LM} \cdot SLA \quad (6.1)$$

where the LAI is measured as m² of leaves per m² of soil; M_{leaf} is the mass of leaves on the 20 plants (g); M_{stem} is the mass of the stems of the plants (g); M_{dm} is the mass of dead material on the 20 plants (g); M_{LM} is the live biomass (kg m⁻²) in each sample plot and SLA is the specific leaf area (m² kg⁻¹) (Roberts et al., 1996; Waterloo et al., 2002).

The vegetation cover and height were obtained from measurements along a transect of 100 m. At each meter, the height was measured with and the percentage of foliar cover was estimated through based on a guide form that was compared to the measurement spot.

The measurements of the ground water depth were made in a piezometer of 50 mm diameter with its screen at 7 m depth. These measurements were taken on a weekly basis and covered the temporal variation in groundwater level well.

Meteorological data were obtained from a micro-meteorological tower, assembled near the biomass sampling site. Radiation was measured by Kipp and Zonen radiometers (CM21 and CG1) (incident and reflecting short wave) from which the albedo was calculated based on the measurements between 10:00 h and 14:00 h (local time) over the period February 1999 until January 2005. Rainfall was measured with a pluviometer (ARG-100). A more detailed description of these measurements is given by von Randow et al. (2004).

6.4 Results and Discussion

6.4.1 Characteristics of the vegetation

The results of this study are presented in Table 6.1 over the period from February 1999 until January 2005. Results obtained from other studies at FNS and at the pasture in RDM are included for comparison.

Table 6.1: Average values and standard deviation of the dead material (kg ha^{-1}), living and total biomass (kg ha^{-1}), SLA ($\text{m}^2 \text{kg}^{-1}$), LAI, coverage (%) and pasture height, other species (kg ha^{-1}) and depth of the ground water in the soil (m) for the FNS during the period from January 1999 to January 2005, compared with the data from other studies in the region.

	<u>FNS</u> <u>Average</u>	<u>FNS</u> <u>Std</u>	<u>RDM</u> <u>Average</u> ¹	<u>RDM</u> <u>Std</u> ¹
Dead material (kg ha^{-1})	2194	888	*	*
Live biomass (kg ha^{-1})	2905	767	2823	*
Total biomass (kg ha^{-1})	5195	1315	2823	*
SLA ($\text{m}^2 \text{kg}^{-1}$)	18.4	5	18	*
SLA ($\text{m}^2 \text{kg}^{-1}$) <i>Panicum maximum</i>	*	*	2.4	*
LAI ^{FNS}	2.0	0.79	1.5	0.2
LAI ^{FNS_Robertset al. (1996)}	2.4	1	*	*
LAI ^{FNS_CarruzoandRocha (2001)}	2.6	0.77	*	*
Cover (%)	80	10.5	*	*
Height, <i>Urochloa brizantha</i> (m)	0,15	0.04	0.31	*
Another species (kg ha^{-1})	90	120	818	*
Groundwater depth (m)	-5.4	0.94	*	*

¹ = Results from Waterloo et al. (2002)

* = No data available

LAI^{FNS_Robertset al. (1996)} = Destructive method (1992–1993)

LAI^{FNS_CarruzoandRocha (2001)} = Optical method from January and February, 1999

The long-term average LAI at the FNS was 2.0. Carruzo and Rocha (2001) observed a higher value using the indirect method (fish-eye, includes grass

stems) in the wet season, whereas Roberts et al. (1996), by using the direct method (which was used in this study), observed higher LAI values (2.5). A reasonable explanation for the LAI differences in relation to the one estimated in this study could be that the sampling period in 1992 and 1993 followed on the planting of the pasture and application of fertilizers with minor grazing activity.

As for the data from RDM, the difference can be due to the composition of the pasture which, in contrast to FNS, had a lower fraction of *Urochloa brizantha* (76%) in RDM, with the remainder being *Urochloa humidicula*, bushes and palm trees. This pasture was previously located in an area of the municipality, which was invaded by small producers and was never cleared from bushes and palm trees, before the land was abandoned. The grass invaded and regrowth resulted in an incomplete cover, which reduced the LAI.

The variation in percentages of live material, stems and dead material to the total was of 29–53%, 37–46% and 12–24%, respectively. An annual average of 2,905 kg ha⁻¹ y⁻¹ was obtained for the living biomass, whereas the dead organic matter reached 2,194 kg ha⁻¹ y⁻¹ for the *Urochloa brizantha* (Fig. 6.1). For the *Urochloa humidicula*, these fractions were of 56%, 23% and 13% for the living material, stems and dead material, respectively, and as much as 7% of plumes with seeds, resulting in an average total value of 90 kg ha⁻¹ y⁻¹.

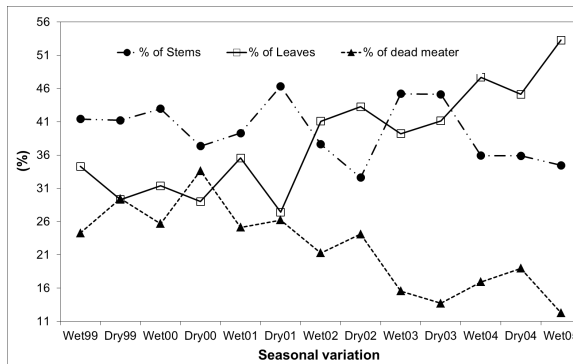


Figure 6.1: Seasonal variation of the percentage of leaves, stems and dead material of the 20 plants for the period of January 1999 to January 2005 at FNS.

The *Urochloa brizantha* SLA was similar at both sites, but lower than that of the *Urochloa humidicula*. Because the biomass and LAI of *Urochloa humidicula* were low, the *Urochloa brizantha* determined the LAI at both sites. We noticed that over the years, there has been a decreasing fraction of dead material and an increasing of the leaf fraction, whereas the fraction of stems remained constant.

The pattern shown in Fig. 6.2 suggests a seasonal variation of dead mate-

rial compared to the total biomass, showing an opposite change over the whole measured period. There is a clear seasonal variation in dead material biomass, which increases in the dry season when the plants become more susceptible to water stress and loss of leaves. Aduan (1998); Baruch and Fernández (1993), reported that in the dry season the plants drastically reduce their surface conductance in response to the higher vapour pressure deficit. Therefore, the dead material showed a strong seasonal variation, directly influencing the seasonal variation in total biomass (Fig. 6.2). On the other hand, the live material also presented a well-defined seasonal cycle in response to seasonal climate variations and the corresponding variation in the soil water availability.

The pattern showed in Fig. 6.2 suggests that the intensity of cattle grazing can affect pasture growth. This is illustrated by the rainy season of 2003, when there was a large reduction in the grazing activity, causing the live biomass to increase from 2,500 to 3,700 (kg ha^{-1}).

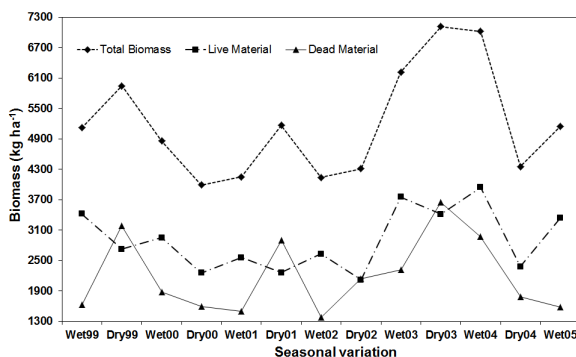


Figure 6.2: Seasonal variation of the dead and living material (kg ha^{-1}) compared with the total biomass (kg ha^{-1}) for all the data collecting period at FNS.

In contrast to seasonal pattern of dead material, the live biomass had a less apparent seasonal pattern to the total biomass. This was due to the well-defined seasonal cycles of the dead material. According to Pereira et al. (1966), the biological responses that influence the adaptation, survival and growth of animals and plants are direct functions of the environment physical variations (*e.g.* temperature, light, water availability and nutrients). Thus, the vegetation possibly responds directly to the event of drought, reducing its aerial biomass and increasing the dead material on the soil.

A well-defined pattern of precipitation input and groundwater level was observed during the analyses made from February 1999 to February 2005. The annual precipitation varied from 1773 to 2295 mm (Fig. 6.3). Both, precipitation and groundwater level depth showed similar seasonal patterns. The groundwater depth reached a minimum at the end of the wet season and then

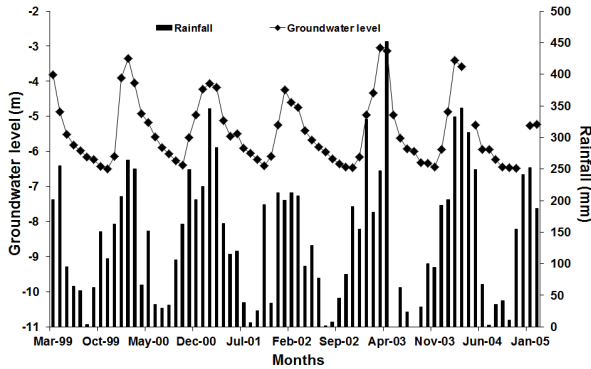


Figure 6.3: Monthly rainfall time series and groundwater depth variation between January 1999 and January 2005 at FNS.

decreased in April and May by about 1.5 m. In addition to drainage, the decrease may be due to the removal of water by plants and evaporation from the wet soil, which was exposed to intense radiation due to the low vegetation coverage, whereas deeper soil moisture would be removed by plant uptake down to rooting depth at over 3 m. Drainage further decreased the groundwater level to 6.5 m below the surface in June (Fig. 6.3).

The vegetation cover clearly responded to the variation in precipitation and groundwater level depth, which is manifested in the seasonal pattern of the LAI. In Fig. 6.4 the wet and dry season average LAI values of the FNS pasture changed in response to variations in soil moisture and groundwater depth. This happens more intensely during the dry period, when the vegetation closes its stomata, thus avoiding the loss of water to the environment. Eagles (1971), Rolim (1980) and Villa Nova et al. (1999) reported several factors that determine the seasonality of forage production, among which the air temperature, solar radiation, photoperiod and soil moisture are considered the main reasons for the physiological vegetation modification.

In the dry season when the soil moisture becomes limiting, and access to the deep groundwater becomes more difficult, plants produce less biomass. The change in the pasture LAI is directly influenced by the seasonal fluctuation of water level in the soil. Thus the seasonal variation of LAI corresponded to the variation of living biomass (Student's *t*-test with 99.9% confidence) and also suggested significantly higher values during the wet season (2.2), than during the dry season (1.7). Monthly LAI values ranged from 0.75 (August 2000) after the cattle grazing in the dry season to 4.1 (March 2004) without grazing.

The pattern of grass height shown in Fig. 6.5 suggested an abrupt decrease of *Urochloa brizantha* height in the wet period for the years 2001 and 2002, to heights similar to those observed during the dry season, thus reinforcing a

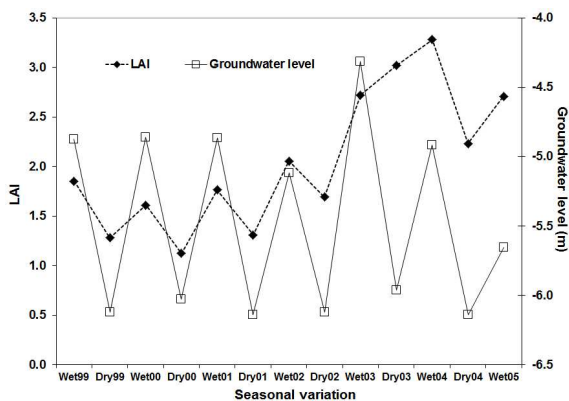


Figure 6.4: Average seasonal pattern of the groundwater level (m) from January 1999 to January 2005 in FNS and corresponding variation of the average seasonal LAI values.

link with the extended period in which the cattle remained in the area in combination with lower precipitation in the dry period from 2001 to 2002. Pereira et al. (1966), through their evaluation of a Brazilian *Cerrado* environment, noted that the performance of ten forage types of grass, with irrigation done in the winter, produced an increase in dry matter of 70% over the studied pasture. On the other hand, in the dry season of 2003, the pasture was higher compared to all other years, which was caused by the absence of intense grazing during a period of low rainfall in the previous months, because the groundwater was lower than in previous years. Moreover, in the beginning of the rainy season of 2003, the precipitation was higher (see Figures 6.1 and 6.4) and the grass height reached to about 0.34 m by the end of the rainy season. During this period the climatic processes combined with a low grazing intensity to provide a dense vegetation cover. The low grazing intensity was due to a change in pasture management to ensure sufficient forage for the ensuing dry period. However, Rodrigues (2004), reports that *Urochloa brizantha* pasture had a rapid growth. In about 28 days of mass production, the pasture could reach from 3 to 6 tons of dry matter per ha along the year. This high production may only be possible under well-watered conditions, low grazing intensity and a high soil fertility. The LAI in this period increased rapidly from 1.5 to 3.0.

The grass height data also showed a decrease over the dry season from June 2004 onwards, but until August of that year there were no cattle in the area and the biomass and LAI was only influenced by climatic conditions. In September 2004, the pasture height showed a strong decrease and remained at 0.14 m in the wet season until the end of January 2005, which showed the impact of a constant presence of cattle in the pasture area.

Results obtained by Roberts et al. (1996) in 1992–1993 for FNS, also based on desstructive sampling, were similar to those presented in this study. Roberts

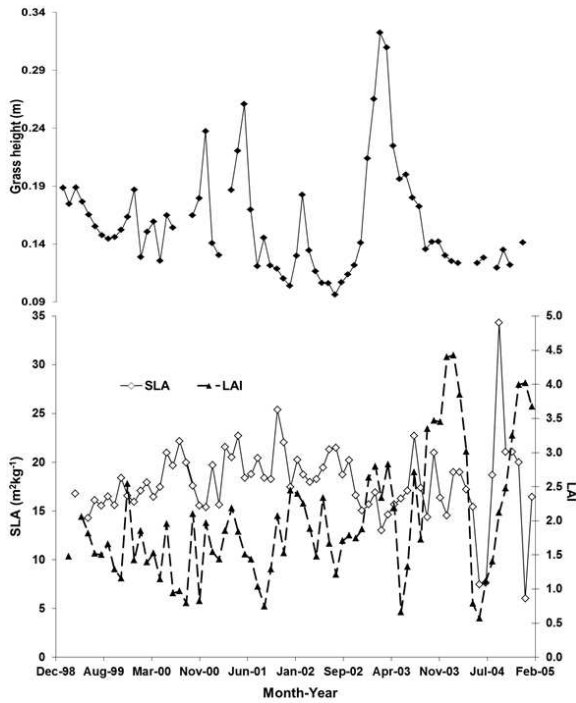


Figure 6.5: Monthly average of the grass height (m), together with specific leaf area ($\text{m}^2 \text{kg}^{-1}$) and leaf area index for the period January 1999 to January 2005.

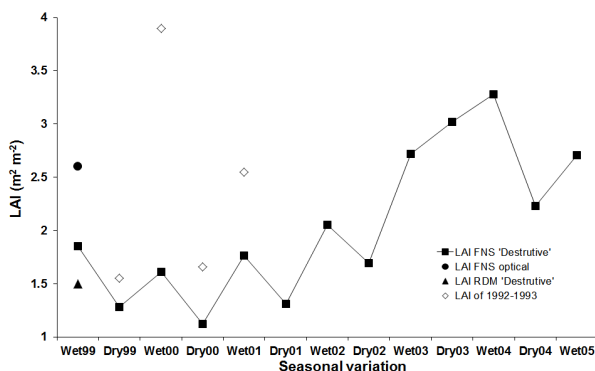


Figure 6.6: Seasonal variation of LAI at FNS. Values of other studies have been included for comparison. Note that these include both destructive and optical sampling methods, and sampling in different years.

et al. (1996) included *Urochloa brizantha* stems in their calculation of LAI, which increased the LAI by about 16–19%. In the dry season, in August 1992, the LAI was 1.55 ± 0.72 , in September 1992, 1.66 ± 0.52 , whereas in the rainy season, April 1993, the LAI was of 3.90 ± 1.02 and in June 1993, the LAI was of 2.55 ± 0.78 (Fig. 6.6). Carruzo and Rocha (2001) estimated the LAI at FNS using the optical method and got an LAI of 2.6 in the wet period. Waterloo et al. (2002) sampled the pasture in the county of Rolim de Moura using eight samples in February 1999, they obtained an average LAI of 1.5. This value was close to the FNS values, for the same period in five samples, in which the average LAI was equal to 1.8, suggesting a similar result for both periods, being the pasture mainly composed by 76% of *Urochloa brizantha*.

The largest differences in biomass were obtained at the end of the rainy season (March and April) and late in the dry season (September and October). The largest variation during the rainy season was in the difference in the quantity of biomass and pasture height in 2003 (no grazing), in comparison to those in the wet season in other years when grazing occurred (see Fig. 6.2 and 6.5). In the dry season, a high of biomass was observed in 2001 and an extremely low amount in 2002 (Fig. 6.7). Not surprisingly, the same extremes were observed in the data of the 20 plant samples and in the LAI (Fig. 6.8).

The conversion of forest to pasture results in a new vegetation cover with low carbon stock, which also impacts the carbon stocks in the soil. The losses in the soil are slower than the loss of carbon during removal of the above-ground biomass, as these occur in small amounts over time (Davidson et al., 1993). Houghton et al. (1987) report that the conversion of tropical forest to pasture would result in a loss of about 25% of carbon stocks from the first 1 m depth. Similar results were found by Eden et al. (1991), reporting a 50% decline in carbon concentration in the pasture soil after 11 years of use in the *Suiá*

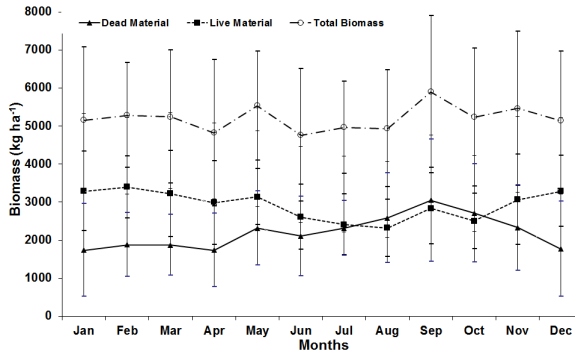


Figure 6.7: Monthly average and maximum and minimum values of living, dead and total biomass ($\text{m}^2 \text{kg}^{-1}$) during the period from January 1999 to January 2005 at FNS.

Missu Farm, Mato Grosso state (Brazil) and about 15% decline in 12 years of pasture usage in an island in Roraima state (Brazil). Moreover, Moraes et al. (1996); Neill et al. (1996) reported that, in Rondônia state (Brazil), in the pasture at *Nova Vida* Ranch (which uses the same pasture rotation technique) the carbon storage in the top 30 cm of the soil increased during their long-term measurements. Fig. 6.9 shows seasonal variation in response to precipitation, but the greatest influence on biomass, height and LAI was exerted by the management in the form of the applied grazing intensity. Fearnside (1989) reported that, in this same pasture, measures of soil carbon productivity indicated that *Urochloa brizantha* pasture produced in 12 years half of the weight of its annual dry material, when compared with a pasture of three years long.

In Fig. 6.9 we observed no for long-term increases or decreases in biomass during the sample period, as could be caused by fertilization or crop management changes. The LAI obtained for the sampling period of 1992 by Roberts et al. (1996) was significantly higher than the current long-term average, presumably due to the lower grazing intensity at that time. However, biomass production may also be decreasing in response to a decrease in soil fertility, perhaps in combination with soil compaction by cattle trampling and exposure to direct action of rain and sun, as suggested by Fearnside (1989).

6.4.2 Pasture LAI and albedo

The albedo of the land surface influences net radiation, and therefore the energy available for evapotranspiration. Fig. 6.10 shows the albedo values that were observed for different LAI values. The albedo showed an inverse relation with the LAI, decreasing from about 0.21 at in the dry season with LAI values less than 1.0 to about 0.17 at LAI values above 3.0. The albedo there-

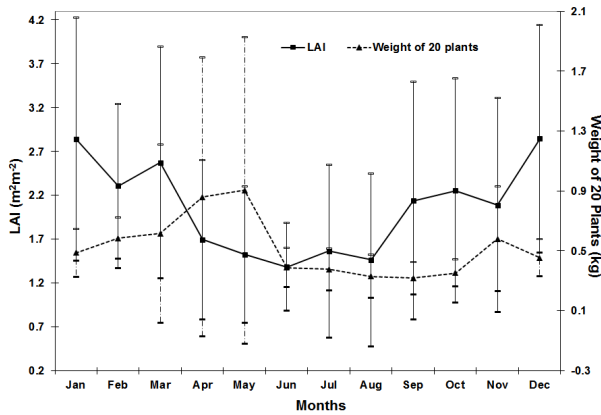


Figure 6.8: Monthly average, LAI maximum/minimum values and the weight of the 20 plants (kg) randomly sampled from each demarcated measured area.

fore shows a similar seasonal pattern as the live grass biomass, and is also influenced by the grazing intensity. The lowest values are observed during the wet season, which means that in this season when soil moisture is not limiting, more energy will be available for evapotranspiration than during the dry season.

In the literature, the albedo value for bare soil is around 0.30. However, conditions in the pasture are never such that the grass dies off completely and the pasture albedo remains in the albedo range for grassland (Waterloo, 1994).

6.5 Conclusions

This study presents the first long-term measurements of biomass and LAI of cattle pasture in Rondônia. The biomass, grass height and LAI showed regular seasonal patterns in response to rainfall during the five years of measurement, which were affected by grazing intensity. With a grazing intensity of 3 heads per ha, the grass biomass decreased to a minimum, with a height of 0.14 m, in contrast to a maximum height of 0.34. The total biomass averaged to approximately 5500 (kg ha⁻¹) in the period from 1999 to 2004, with the largest biomass reached in 2003, approximately 6300 (kg ha⁻¹) when no cattle grazing occurred. There was no evidence that the biomass production rate changed over the 5 years of measurement, as shown in Fig. 6.9.

The long-term average LAI (2.0) was somewhat lower than the values observed in campaigns in the past by Roberts et al. (1996, 2.6) and by Carruzo and Rocha (2001, 2.4). Roberts et al. (1996) recommend the inclusion of the stem in the LAI, which would result in a closer match with our estimates. On the other hand pasture conditions in 1992 were different as the grass had re-

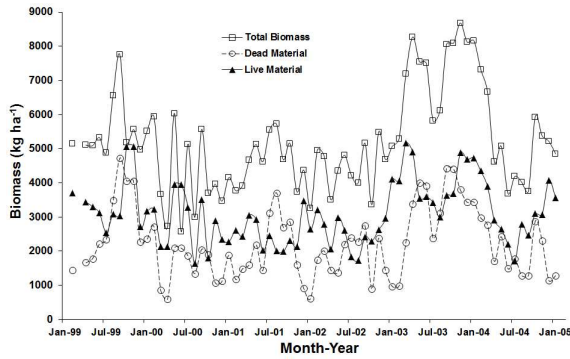


Figure 6.9: Monthly averages of living biomass, dead plant material and total biomass for the period from February 1999 until January 2005 at FNS.

cently been planted and there was no grazing. Inclusion of stems in the LAI calculations would increase the LAI values by 16–19%, from 1.7 to 2.0, thus increasing the LAI calculation estimates at the FNS, thereby approximating the results obtained using the optical method by Carruzo and Rocha (2001). Monthly LAI values varied between 0.5 in the dry season with grazing to 4.1 in the wet season in the absence of grazing.

The albedo showed an inverse relation with the LAI and therefore also responds to changes in rainfall and grazing intensities.

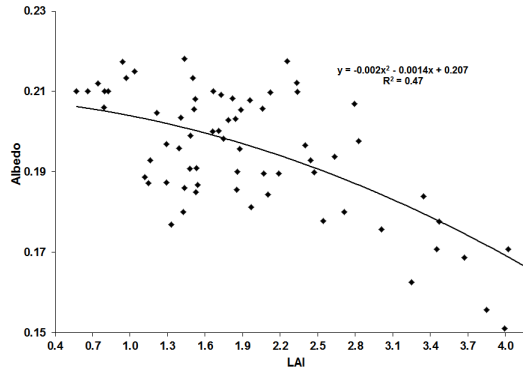


Figure 6.10: Correlation of pasture LAI and the average albedo (10:00–14:00 h) for monthly values measured from January 1999 to January 2005 at FNS.

Chapter 7

Synthesis and recommendations

7.1 Introduction

Several studies have shown that the interaction of the Amazon rainforest with climate, climate change and human pressure can exert a significant impact on the vulnerability of the Amazonian landscape (Costa and Foley, 2000; Nobre and Borma, 2009; Lewis et al., 2011; Aragão, 2012; Spracklen et al., 2012). In particular, Earth System Models (ESMs) (Oyama and Nobre, 2003; Cox et al., 2004) suggest a vulnerability of the vegetation to future changes in precipitation regimes that may further change with global climate change. If predictions of drier and warmer future climatic conditions hold true, even larger portions of Amazonia will soon be vulnerable to fire (Le Page et al., 2010). Most climate models, but not all, predict substantial reductions in precipitation by the mid 21st century in eastern Amazonia due mostly to increased atmospheric CO₂ (Malhi et al., 2008). By increasing the fire frequency and intensity, this change in climate is anticipated to have substantial effects on tropical forest dynamics (Barlow and Peres, 2008; Nepstad et al., 2008).

Most of the central Amazonia landscape consists of toposequences from plateau to valley forest (Fig. 1.3). Whilst Oxisols and Ultisols on the slopes and plateau have a significant clay fraction, valley soils generally consist of strongly leached quartz sands (Spodosols) with little capacity to retain water or nutrients. These toposequences generally have a direct groundwater connection between them through recharge (Hodnett et al., 1997; Waterloo et al., 2006; Tomasella et al., 2007). Permanently wet soils in the valleys are maintained by a continuous supply of groundwater from the plateau and slope areas. Thus, there is the possibility that any changes in the water budget of the plateau areas would also affect the functioning of the forests in the valley areas. Because the Central Amazonian area is very inhomogeneous, consisting

of a mosaic of well-drained dissected plateaus, separated by steep slopes and sometimes broad swampy river valleys (Waterloo et al., 2006; Nobre et al., 2011), this can potentially affect a large area of central Amazonia.

To understand the vulnerability of the Amazon rainforest to a long-term decrease in rainfall (Oyama and Nobre, 2003; Nobre and Borma, 2009). Nepstad et al. (2002b), Meir et al. (2009) and Costa et al. (2010) conducted artificial rainfall exclusion experiments (e.g., SECAFLORESTA and ESECAFLOR). These studies demonstrated that drought-stressed trees develop a physiological adaptation, which maintains a relatively constant water tension in the xylem in both wet and dry seasons making them resilient to short term droughts. With progressing long-term drought, eventually mortality kicks in, with the roots becoming unable to extract enough soil water (Fisher et al., 2006; Costa et al., 2010). Mortality rates in the drought plots increased to nearly three times that in the control plot during years 4-7 of partial rainfall exclusion (Meir et al., 2009).

Similar in concept, in our study, we performed a drainage experiment, by installing a set of drainage tubes at about 1 meter depth, thus artificially enforcing a lowering of the groundwater table, that was previously located at the surface or less than 30 cm below it (see Fig. 5.2). The experiment mimicked a possible rainfall decreases in the toposequence area (plateau-valley) of central Amazonia. Because the climate in this region shows little seasonality (Satyamurty et al., 2010). Overall, it aimed to identify the sensitivity of the Amazon valley forest to rainfall reduction, as one of the main factors determining the forest resilience to climate change. In a more general sense, the present study aimed at finding patterns of change in water and nutrient cycling resulting from this experiment and from the experimental monitoring of an undisturbed control plot next to the manipulated area and seven other Amazonian sites at Cuieiras Reserve (plateau, slope, *Campinarana*, valley forests) and in the Campina Reserve (tall heath forest-*THF*, stunted heath forest-*SHF* and bare soil areas). We now discuss our main findings.

7.2 Soil CO₂ emission

We characterized the soil CO₂ emission by improving the methodology to measure in the sandy and loamy soils of the rain forests (Chapter 2). We noticed a strong effect of rainfall and the insertion of the rings used to fix the chambers. The ring insertion caused an increase of about 13–20% in the mean value of soil CO₂ efflux for the first 7 hours. When the rainfall was between 3 to 8 mm, we noted an increase in the emission of about 10 to 18% for the next 2 hours, while for larger storms we noted a decrease due to saturation of the pores with infiltrated water that could increase 27% of the original value. The new methodology allowed us to calculate the total emission of the area $3.08 \pm 0.8 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ for the Campina Reserve and at $3.82 \pm 0.76 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ for the Cuieiras Reserve. In chapter 3 we determined the

relationship between soil respiration and soil temperature. A new analytical model that was able to take into account the hysteresis between the amplitude/phase of the respiration and the soil temperatures at various depths improved previous model estimates. The analysis also showed that the largest time lag between soil respiration and soil temperature occurs in forests on poor soils while the smallest time lags were found in tall forest. This suggests that a relation between the LAI of the forests and the time lag in soil respiration may exist, with smaller time lags for high LAI forests, and increasing time lags for forest with lower LAI. The latter consist of an increasing number of more scleromorphous leaves that are associated with availability of nutrients. The occurrence of scleromorphic vegetation is known to be caused by a slow translocation mechanism of photosynthate to the roots and associated microbes as reported for species with slow growth rate (Baldocchi et al., 2006; Metcalfe et al., 2011).

7.3 Biogeochemistry of a forested basin

Another important result arises from the implementation of the first integrated study of organic/inorganic carbon, nutrient and water balance export of the *Campina* forest (Stunted Heath Forest–*SHF* and Tall Heath Forest–*THF*) in central Amazonia (chapter 4). The results from these *Campina* forests suggest that a drier climate with falling groundwater levels in the valleys (See Chapter 5) would result in higher DOC and lower DIC concentrations in groundwater, thus affecting the baseflow concentrations of these rainforest catchments. The nutrient data suggested that the *Campina* catchment was somewhat less efficient in retaining K and Ca, which can be explained by the low nutrient retention capacity of the *Campina* sandy soil. The *Campina* forests transpired much less than the *Terra-Firme* and also showed less overall evaporation. The runoff coefficient for the *Campina* catchment also indicated lower evaporation in comparison to that of *Terra-Firme*. The *Campina* forest also had higher albedo compared to the tall evergreen forest thus having less energy available for evaporation.

7.4 Responses to drought

If precipitation drops below 1800 mm year⁻¹ in Amazonia, sustained a longer dry periods, it will may lead to vegetation loss and eventually to replacement of rainforest by other types of forest. The transition path between these vegetation types and associated climate states determines the resilience and stability of the tropical carbon reservoir. Several factors combine to determine the exact boundaries in space and time between valley forests and other types such as *Campinarama*, *Campina* and ultimately savannah. However long-term biomass consumption and mortality due to fires or drought may ultimately

favour the occurrence of new other types in place of valley forests if precipitation declines. Thus, changes in mortality due to increases or decreases in drought duration and fire frequencies play a critical role in the shift of one climate-vegetation state to the other. We tried to mimic this potential trajectory by implementing a drainage experiment. We do note that this was a sudden rather than a gradual change. The main result of the artificial drainage experiment (Chapter 5), was a change in the loading of dissolved organic carbon (DOC), soil respiration, LAI, precipitation, ecosystem production and soil temperature (Fig. 7.1) compared to all other ecosystems we studied.

The results of the valley forest were obtained for a relatively short period of time that reduced the groundwater recharge but showed that the sandy soils rapidly lost their soil organic material and nutrients. The soil CO₂ respiration also decreased to a similar rate as that of the bare soil emission from the stunted heath forest (*SHT*) (chapter 5). Soil carbon, soil nutrients, dissolved organic/inorganic carbon (DOC/DIC) and the soil organic matter (SOM) content, also decreased to a similar level as that of the *Campina* forest (chapter 4). The quartz sand formation of the valley forest combined with artificial drainage helped to increase the carbon and nutrient weathering to be either flushed away to the lower groundwater or to exfiltrate to another area, forcing the valley forest soil ultimately to be similar to the *Campina* forest.

7.5 Synthesis, multiple stable states

We started this study realizing the existence of a lack of knowledge about the biogeochemical functioning of central Amazonian toposequences, the transect from plateau to valley forest. Lacking this knowledge it was impossible to determine whether the overall system was resilient to seasonal and moderate drought or to other disturbances. Repeated or prolonged disturbances have the potential to change the forest structure (land-cover) and nutrient dynamics. This can lead potentially to long-term changes in vegetation composition and associated C loss (Nepstad et al., 2002b; Aragão et al., 2008). The severe droughts of 2005 and 2010 provide an early warning to the possible consequences. As these droughts are related to subtle changes in the Atlantic Multidecadal Oscillation and El Niño/Southern Oscillation (ENSO), they may become more prevalent in future in the Amazon region (Marengo, 2004; Nepstad et al., 2008). The anomalous warming of the tropical North Atlantic is generally identified as the cause of the 2005 drought. Cox et al. (2008), analysed GCMs for the Atlantic N-S gradient (an index of the north-south SST gradient across the equatorial Atlantic) that is associated to a reduction of dry-season (July-October) rainfall in western Amazonia and suggested an increased frequency of severe droughts. The severe drought in 2005 in south-western Amazonia with dry season temperatures of 3-5 °C warmer than normal and rainfall over the Rio Solimões basin at only 33-65% of average values, showed the dominance of mortality over “simple” GPP drought effects. Analysis of the

long-term biomass monitoring plots across Amazonia furthermore showed that most of the forest plots experienced severe water deficits in 2005 losing several tons of living tree biomass carbon per hectare (Phillips et al., 2009). Human pressure and frequent fire in the forest may also change the structure, composition and functioning of vegetation by selecting fire adapted species and favouring more flammable species (for example, grasses) and this may lead in specific areas to more savanna-like ecosystems (Soares-Filho et al., 2004; Nepstad et al., 2008; Davidson et al., 2012). Ultimately, the environment determines selectively which ecological strategy is successful. The combined life histories of species have thus ensured the persistence of multiple species across the landscape in accordance with resource gradients and disturbance regimes (Bonan, 2008b). Savannization shifts can then be regarded as simply one of the Amazonian vulnerability options.

The hypothetical ecosystems perturbations of Hirota et al. (2011) are based on the response to gradually changing conditions of annual precipitation (rather than our sudden abrupt change in the drainage experiment), together with climate change, nutrient charging and habitat fragmentation or biotic exploitation in the Amazon (Nepstad et al., 2008; Davidson et al., 2012). Although diverse events can trigger such ecosystem changes, recent ecosystem studies (Soares-Filho et al., 2004; Nepstad et al., 2008; Davidson et al., 2012) show that low ecosystem resilience usually paves the way to an alternative state. This suggests that strategies for sustainable management of such ecosystems should focus on maintaining resilience, avoiding the *Business-as-usual* scenario that would reduce about 40% of the Amazon basin forest (Fig. 1.1).

While it is impossible to address all the relevant issues in these complex shift in one thesis, we can use some of our data to further investigate the sensitivity of our observational transect to changes in key indicators such as soil emissions, DOC transport and LAI. The data in Fig. 7.1 consist of data from the Chapters 3, 4, 5 and the authors Salimon et al. (2004), Pinto-Junior et al. (2009) and Marques (2009). The figure indicates a tendency, not more, of how a valley forest ecosystem could gradually turn into *Campina* forests by developments similar to those observed in our artificial drainage experiment. It can be noted from this Figure that the LAI, as an indicator of carbon storage, declines along the trajectory.

Furthermore, intensive and prolonged droughts can easily lead to very poor and potentially toxic soils, stimulating a potential change from the existing tall valley forest to increasingly a tall/stunted heath forest (THF/SHT) with more sclerophyllus leaves. This would be the *Campinarana* or, once the decline goes further, the even more dry forest the *Campina*. Because climatically induced drought conditions may cause degradation of wet valleys and turn them into *Campina*-like forest, the LAI eventually attains similar levels as that of pasture (Chapter 6). Low LAI then results in less evapotranspiration and possibly less carbon and water recycling as observed in the current study.

The early modelling studies (Cox et al., 2000; Oyama and Nobre, 2003; Cox et al., 2004), indicated that Amazon savannisation had a high likelihood,

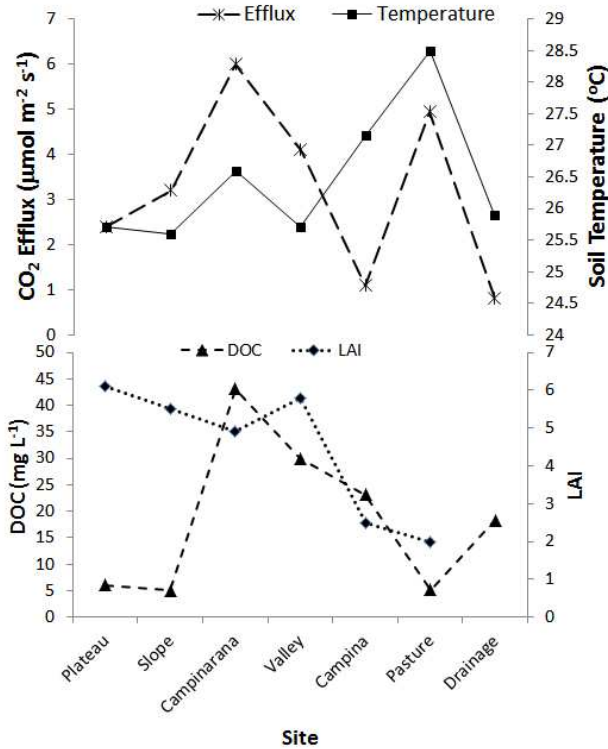


Figure 7.1: Effect on the resilience of multi-stable ecosystems using DOC, LAI, soil respiration and temperature. The figure indicates the tendency of the indicators for a vegetation shift from the tall evergreen forest to the drainage forest, that has similar behaviour as the well-drained open forest of *Campina*.

as a result of declines in precipitation and increases in temperature. More recent studies suggest that the specific climate models used may have amplified the tendency to strengthen the positive feedback between declining rainfall and vegetation in the model (Cox et al., 2013; Huntingford et al., 2013). These recent studies suggest that the range over which the forest declines may be in fact thus much smaller. Ultimately the question whether a dieback may occur or not, depends critically on the balance between changes in climate ecosystem, water and nutrient balance, future land-use change and, importantly, how increased CO₂ concentrations affect the net ecosystem uptake of carbon (the CO₂ fertilization effect) (Huntingford et al., 2013). The impact of the CO₂ fertilization effect is unfortunately very poorly known for tropical rainforest. As such, our findings from the *Campina* forest together with the results from

the artificial experiment in the valley, are important to provide insights into possible changes in evapotranspiration, soil water chemistry, carbon and inorganic nutrient export from the valleys in the central Amazonian rainforest. They show potential trajectories for changes under a drier climate.

7.6 Recommendations

Our work has comprised the application of a number of experimental methods in a harsh environment. There are several lessons to be learned from this work. For future studies, we recommend replication of the artificial drainage experiment (Chapter 5) in different valley areas over the Amazon region, so that a statistically sound analysis can be performed, and Fig. 7.1 could be further substantiated with more results. With a similar methodology, but in different areas around the Amazon, it would be possible to more realistically assess the effect of decreasing rainfall effect on the central Amazon rain forest landscape, using more replication than with our single pilot drought experiment.

The experiments did generate a set of interesting results for the valley forest, but was also a serious challenge to operate because of the high rainfall in the Amazon and the partly swampy valley soils. The results of the artificial drainage experiment are however only confined to areas with similar topography and toposquence of plateau-slope-valley forests in Central Amazonia. The advantage compared to the ESECAFLOR and SECAFLORESTA experiments was that the experiment does not need very large maintenance to exclude the rainfall from the area. The big disadvantage was the disturbance of the soil as a result of the digging of three soil trenches which destroyed the fine root system, but maintained the larger roots. The major maintenance in the drainage experiment was to unlock the output of the drainage tubes once they were blocked with leaves. Another important advantage was the fast change in soil characteristics, but this demanded also a high measurements frequency. In our case, the first six months were crucial to identify the drought sensitivity and to observe the major soil carbon and nutrient and soil CO₂ evasion decreasing during the initial period. Future experiments would thus do well to be repeated on several spots and be monitored for a much longer period.

We were able to estimate the water balance and the carbon and nutrient export from the *Campina* catchment, but for future work we suggest a more precise determination of the water balance by using eddy covariance measurements together with sap flow, interception, baseflow and stemflow methods. This would minimize the uncertainty in the evapotranspiration estimates in relation to the edge of the forests (THF-SHF). Further studies are needed to understand the precise biogeochemical processes that lead to DOC formation in the well-drained *Campina* forest soils and in the poorly drained valley soils of the rainforest catchments.

In addition, we recommend increasing the number of soil respiration systems to measure simultaneously in order to obtain a reliable data set from the

area. It is also important to include soil tensiometer measurements next to the soil chambers, as this would give support to further validate the analytical solutions on soil water matrix potential in the soil respiration models for the Amazonian ecosystems. It will however remain a challenge in the Amazon region to find a well-defined response of soil respiration to the standard soil temperature and/or soil moisture observations, as the range in temperature and moisture will always be small. Therefore, we spent considerable effort on improving our soil respiration methodology. Using this we were able to use the time lag as an indicator of forest function and characterize the ecosystem behaviour in terms of the soil respiration and forest and soil type. New studies on soil respiration could benefit from this methodology to further understand the complex interrelations between soil, water and vegetation biogeochemistry that ultimately determine the resilience of the Amazonian rainforest system.

Chapter 8

Summary

The PhD study titled *Vulnerability to drought and soil carbon exchange of valley forest in Central Amazonia (Brazil)* investigated the potential changes in the water and carbon dynamics of central Amazonian rain forest in response to a potential future change in climate leading to reduced rainfall inputs. A drier climate might change the water-logged condition of sandy soils in valleys, which might induce a change in forest composition towards that of low-statured campina forest. This study was performed in the Cuieiras and Campina Forest Reserves in Amazonas State and in a cattle pasture in Rondônia State in Brazil. The study sites cover seven distinctly different topographic, soil and land cover environments, *i.e.* pasture in deforested area in Rondônia, stunted and tall heath forests on sandy soils in the Campina Reserve and a topo-sequence of riparian and ecotone campinarana forests on sandy soils and tall *terra firme* forests on slopes and plateaus with clayey soils in the Cuieiras Reserve.

The objectives of this PhD study were focussed towards:

- studying soil respiration, and model surface temperature dependence of soil respiration for different land cover and soil types;
- quantification of the water balance and carbon and nutrient exports of campina heath forest for comparison to *terra firme* rain forest;
- evaluation of potential climate change impacts (drought) by comparing soil nutrient and carbon contents and soil respiration in an artificially drained sandy valley riparian forest soil with values observed for campina forest soils, and;
- quantification of the long-term variation in biomass and leaf cover in a seasonal cattle pasture and comparison with LAI values measured in the various rain forest types.

The scientific knowledge gained in this study provides insights into the functioning of different forests types (riparian, campina, campinarana, tall rain forest) that are common in the central Amazon landscape.

8.1 Soil CO₂ respiration

Soil respiration plays a significant role in the carbon cycle of Amazonian rain forests, but measurements have yet only been carried out in few locations. We used automatic soil respiration systems to measure time series of soil CO₂ fluxes and soil temperature at 5 cm depth at 30-minute intervals.

The effects of measurement chamber ring insertion into the soil were quantified to improve the methodology. In addition, the impact of precipitation on soil respiration fluxes and the spatial distribution of CO₂ emission in the central Amazon were studied using measurements made in seven different forest locations.

The ring insertion effect increased soil emission by 13% to 20% for sandy and loamy soils for four to seven hours after insertion, respectively. The loamy soil therefore seems to be more sensitive to mechanical disturbance. This suggests that when measurements are consistently made shortly after ring insertion soil CO₂ emission could be overestimated by as much as 20%.

Soil respiration showed no change following precipitation events below 3 mm. For higher rainfall rates differences in response were observed for loamy and sandy soils. For loamy soils, precipitation events between 3 mm and 8 mm caused an increase of up 18% for two hours after the event, followed by a reduction of 15% for the next 15 hours. Precipitation amounts higher than 8 mm caused an immediate reduction of 22–27% in CO₂ respiration until 18 hours after precipitation had ceased. In sandy soils, the response was much faster but of shorter duration. For precipitation events between 3 and 10 mm a 34% increase in CO₂ respiration was observed within two hours after precipitation ceased, but fluxes returned to pre-event levels during the next two hours. Precipitation above 15 mm resulted in a fast reduction in soil respiration lasting for about 5 hours.

Soil CO₂ respiration varied significantly between the seven different forest environments. The lowest values (0.9–1.2 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) were observed for bare soil and stunted heath forest on the white sandy soils in the Campina Reserve, whereas that for tall heath forest was higher at 3.8 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$. In the Cuieiras Reserve, values for tall rain forests on clayey soils on plateau and slope varied between 2.4 and 3.2 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$, respectively. Soil respiration increased in the sandy soils in the valley under riparian forest (4.1 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) and were highest in the campinarana forest at 6.0 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$.

The soil CO₂ respiration fluxes for the complex forests of the Cuieiras and Campina Reserves were estimated by combining the measurements made in the different forest environments in the reserves with the spatial extent of

these environments as determined from Shuttle Radar Topographic Mission (SRTM) data. For the Cuieiras Reserve, which included plateau, slope, campinarana and riparian forests a respiration of $3.8 \pm 0.8 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ was observed, which was higher than the $3.1 \pm 0.8 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ observed for the mosaic of bare soil, stunted and tall heath forest in the Campina Reserve. Because the present values incorporate spatial variations related to differences in soil type and vegetation cover they may be considered more representative than those published in earlier studies that did not explicitly account for the cover fractions of different forest types. The study shows that for modelling CO₂ respiration of non-homogeneous areas information about the functioning of different landscape units can reduce uncertainties in carbon release.

Automated soil respiration measurements allow a better understanding of the possible factors that drive soil respiration processes. The main factor controlling soil respiration, excluding periods with rainfall, was soil temperature, with 90% of the variation explained by regression analysis. Rainfall seemed to contribute only by causing a quick decrease in soil temperature during percolation and consequently the respiration followed the physical effect of soil water percolation.

Forest leaf area index (LAI) were measured with a Li-Cor 2000 plant canopy analyzer in the different forests. LAI varied between 2.5 and 4.1 in the stunted and tall heath forests in the Campina Reserve, respectively. The LAI was higher in the Cuieiras Reserve, varying between 4.9 in the campinarana forest to 5.5–6.1 for the tall rain forest on slopes and plateaus.

The soil respiration measurements for the seven different forest environments were also analysed to investigate differences between sites and their causes using a modelling approach. Ensemble averages of hourly fluxes were calculated for wet and dry seasons (as far as these were sampled) from the 30-minute interval time series. These values were processed using an analytical model estimating the soil surface temperature from the temperature observed at 5 and 10 cm depths.

Relationships were derived between the amplitude and phase of the soil CO₂ respiration and the soil temperatures at various depths in the soil. Compared to the use of soil temperatures at 5 cm and 10 cm depth, the use of (estimated) surface temperatures strongly reduced the hysteresis between T and R_s and improved the coefficient of determination for most forest sites.

Using a simple exponential model to express soil CO₂ respiration in surface temperature improved the correlation between soil temperature and R_s on a daily time scale. The Q_{10} surface temperature sensitivity ranged from a low 1.3–1.7 for heath forest in the Campina Reserve to a high 7.7 for the wet riparian forest soils in the Cuieiras Reserve. R^2 -values ranged from 0.42–0.92. Q_{10} values for tall rain forest on clayey soil were between 3.8 (slope) and 5.2 (plateau). Differences in forest composition and soil type corresponded to observed differences in temperature time lags, with the longest time lags observed for the slowest growing *Campina* vegetation.

8.2 Water balance quantification and carbon and nutrient export from the Campina forest Reserve

Published water balance studies for the Amazon are almost exclusively based on integrative micro-meteorological and catchment water balance studies for *terra firme* rain forest and do not provide information for specific vegetation types in the region, such as for riparian or heath forests. Similarly, carbon storage values have been studied through different approaches to assess whether the rainforest ecosystem is likely to act as a sink or source for carbon in the near future. A water balance and carbon export study carried out in a micro-scale heath forest (*Campina*) catchment in the Campina Reserve. This study publishes the first water balance component estimates for a stunted and tall heath forest catchment through combination of a catchment water balance study, micro-meteorology and sapflow measurements and provides information on the dissolved carbon and selected nutrient exports.

For a one-year study period (18 March 2007 until 19 March 2008) rainfall amounted to 3054 mm, which was about 800 mm above the long-term average for this region. Corresponding forest evaporation amounted to 1532 mm (4.1 mm day^{-1}), of which 899 mm was evaporated under dry canopy conditions from stunted and tall heath forest. Rainfall interception loss amounted to 15.6% of gross rainfall, assuming a low stemflow fraction based on literature values. Transpiration by the stunted heath forest, estimated from sapflow measurements, was much lower at 475 mm, which amounts to about 60% of that of tall heath forest. Surface runoff amounted to 485 mm, whereas another 1071 mm was discharged as regional groundwater outflow. In years with average rainfall, interception loss will be lower and soil moisture stress may play a role. Hence evapotranspiration for the campina forest may be expected to be lower than that of rain forest growing on soils with higher plant water availability. The higher runoff coefficient observed for the Campina catchment also indicates a lower evaporation in comparison to that by rain forest.

Accumulated dissolved organic and inorganic carbon (DOC and DIC) exports in surface runoff amounted to 15.3 and $3.5 \text{ g m}^{-2} \text{ yr}^{-1}$, respectively, with corresponding daily exports of $41.8 \text{ mg m}^{-2} \text{ day}^{-1}$ and $9.5 \text{ mg m}^{-2} \text{ day}^{-1}$. Groundwater DOC and DIC exports were higher at 31.7 g m^{-2} and 5.5 g m^{-2} , respectively. The total carbon exported in one-year period were 55.9 g m^{-2} . This is much higher than that observed for the catchment in the Cuieiras Reserve (DOC export $<20 \text{ g m}^{-2}$). As campina forest areas cover a significant proportion of the Amazon Basin, these differences in ecosystem carbon exports should be taken into account in future studies assessing the carbon budget for the Amazon Basin.

8.3 Climate change (drought) effect on valley forest soil

Large-scale conversion of rain forest to agricultural land may cause changes in local, as well as global water and carbon cycles. For the Amazon reductions in precipitation has been predicted from model simulations. Changes in the hydrological cycle may potentially lead to dangerous positive feedback effects in the carbon cycle through tree mortality (drought), changes in CO₂ emissions and in the forest CO₂ sink strength, decreased biomass production and reduced evapotranspiration by the remaining forest. Forests on sandy soils cover a significant part of the central Amazonian landscape, approximately 6% of Amazonia (McClain et al., 1997; IBGE, 1997; Luizão et al., 2007). In central Amazon the poorly-drained valley soils are of similar sandy texture as those of soils under campina forest cover, but support riparian rain forest. Under present climate conditions the vegetation in the riparian zone is not likely to experience significant periods of droughts. However, a lowering of ground water in these valleys under drier climate conditions may induce biochemical changes in the soil such that the riparian forest adapted to the poorly drained status cannot be supported any more leading to a change in forest composition.

In this study spatial and temporal variations in litter decomposition, soil and groundwater chemistry and soil CO₂ respiration were studied in forests on sandy soils (riparian, campinarana, campina), whereas the drought sensitivity of poorly-drained valley soils was investigated in a drainage experiment to artificially lower the ground water table.

Depending on the drainage status of the sandy soils, a large natural variation was observed in soil carbon, nitrogen, exchangeable cation concentrations, soil water chemistry and litter decomposition rates. Litter decomposition and soil CO₂ respiration rates were very low in *Campina* forests, where the capacity of the coarse sandy soil to retain nutrients was extremely low. The highest soil CO₂ respiration rates within the *terra firme* landscape were observed in ecotone *Campinarana* forests that form the transition from riparian forests on near saturated sandy soils to tall *terra firme* forests on slopes and plateaus with clayey soils.

Artificial drainage of the riparian forest plot lowering the groundwater level by a modest 0.3 m, did not cause apparent changes in the soil nutrient content, chemical composition of soil water or litter decomposition rates within the study period of one year. However, marked reductions were observed in the top soil carbon (1.4 to 0.1%) and nitrogen (0.05 to 0.00%) concentrations, as well as in the soil CO₂ respiration rate. The latter decreased from 3.6 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ at the start of the drainage experiment to 2.1 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ eight months into the experiment, which is similar to those observed in the campina forest. The drainage study therefore indicated that the riparian forest soil is very sensitive to changes in the groundwater level. It remains unclear if slower changes in soil and water chemistry in the subsoil

would become apparent at a later stage. This suggests that drainage of the sandy riparian forest soils would induce a transition towards a drier campina-type forest, with low soil nutrient contents and low soil respiration. On the other hand, a shift in vegetation species and the development of a root mat in time, such as present in the adjacent ecotone campinarana forest, would result in increased soil respiration rates. It is yet uncertain what the direction of succession of drying riparian forest will be and continued monitoring of the drained and control plots may provide valuable clues to future trends.

The large difference in soil respiration rates observed between *Campina* and *Campinarana* forests imply that knowledge about the likelihood of riparian forest to be succeeded by campina or ecotone campinarana forest is crucial for determining the future carbon budget of central Amazonia *terra firme* rain forest ecosystem under a drier climate.

8.4 Biomass and Leaf Area Index of cattle pasture in Rondônia

The Amazon Basin is covered by a mosaic of undisturbed natural rain forest, secondary growth forests, cattle pastures and agricultural crop lands. In Rondônia State, rain forests has been converted to cattle pasture since the 1950s. In contrast to evergreen rain forest, the cattle pastures can be expected to have a distinct seasonal growth cycle, which would result in significant variation in the leaf cover (LAI) and biomass. LAI is an important land surface parameter for modelling of climate and hydrology, but little is known about its seasonal variation in the Amazon. This long-term study (1999–2005), based on destructive measurements, was the first to describe the seasonality in biomass production and LAI from Amazonian cattle pasture. The grazing intensity was three animals per hectare at maximum and the dominant grass species was *Uruchloa brizantha*.

The live grass biomass averaged 2905 kg ha⁻¹ over the five-year period, whereas the grass litter amounted to 2194 kg ha⁻¹, which results in a total average above-ground biomass of 5195 kg ha⁻¹. The monthly biomass total went up to 8000 kg ha⁻¹ during the wet season of 2004 in the absence of grazing.

The average LAI of the pasture was 2.0. Monthly LAI values varied seasonally, but also with the intensity of grazing. Dry season LAI values were typically around 1.2 when grazing occurred, but remained above 2.0 in the absence of grazing. Average wet season LAI values were only slightly higher with grazing (1.5–2.0), but went up to 3.3 in the absence of grazing. The maximum monthly LAI measured was 4.0. The corresponding pasture albedo varied from about 0.20 for LAI values below 2.0 to 0.17 at LAI values above 3.0. The long-term measurements indicate that the pasture biomass, LAI and albedo show strong influences of seasonality in the absence of grazing, which

is suppressed when grazing occurs. On a monthly basis, the LAI of the cattle pasture can therefore vary between 0.5 and 4.1. These LAI values are comparable to those measured in the Campina Reserve heath forest (2.5–4.1).

8.5 Conclusions and recommendations

This study has shown that significant differences exist in the hydrological functioning and the nutrient and carbon cycling of different forest types that make up the Amazon rain forest biome. The drainage experiment indicates that a lowering of the phreatic level in the valleys will have consequences on the soil carbon emissions, and presumably also on the forest cover if the lowering becomes permanent. It is yet uncertain if this would lead to a change of riparian forest to campina forest (with low soil respiration values and water use) or to campinarana forest (with high soil respiration). Additional knowledge about the functioning of these different forest types, and the impact of climate change (drought) on the future spatial distribution of these forests, should be obtained in long-term studies for further evaluation of the sensitivity of the biome to change and the impacts on the water and carbon cycles.

Chapter 9

Samenvatting

Het promotieonderzoek getiteld *Gevoeligheid voor droogte en bodemkoolstofuitwisseling van het regenwoud in Centraal Amazone (Brazilië)* heeft de potentiële veranderingen ten gevolge van klimaatverandering met afnemende neerslag onderzocht in de water- en koolstofkringlopen van het regenwoud ecosysteem in centraal Amazone. Een klimaat met minder jaarlijkse neerslag kan invloed hebben op de hoge grondwaterspiegel in de zandige bodems in de rivierdalen, wat weer zou kunnen leiden tot een transitie van broekbos in de rivierdalen naar heidewoud (*campina* of *campinarana*) bos.

Het regenwoud in het centraal Amazonegebied is niet homogeen. De soortensamenstelling van het bos, en de hoogte van het kronendak, vertonen aanmerkelijke variaties in relatie tot bodemtype (zand, klei) en de lokale topografie en grondwaterstand (aan maaiveld in de rivierdalen tot 40 m diep onder de plateaus). Campina heidewoud, in het Campina Reservaat, groeit op een zeer arme bodem van uitgeloofd wit kwartzand, waar de grondwaterstand meer dan 2 m diep is. Dit type bos heeft een aanmerkelijk lagere soortenrijkdom dan die van het regenwoud op kleibodems. In het gedrongen campina heidewoud is het kronendak slechts 4–7 m hoog, en is de bodem onbegroeid tussen de groepjes bomen in. In het aanliggende hogere campina heidewoud reikt het kronendak tot 18 m. Campinarana bos, in het Cuieiras Reservaat, groeit ook op uigeloofd kwartzand, maar de grondwaterstand ligt hoger dan in het campina heidewoud, tussen de 0.5–1.0 m onder het oppervlak. Campinarana bos vormt een overgang van broekbos, waar de grondwaterspiegel in de bodem van kwartzand aan, of net onder, het maaiveld staat, en het tropisch regenwoud wat op kleibodems groeit op de hellingen en plateaus in het gebied. Het grondwater onder het tropisch regenwoud staat veel dieper, tot 35 m diep onder het plateau. Het kronendak reikt in broekbos en campinarana bos tot ongeveer 25 m hoogte, terwijl dat van het tropisch regenwoud tussen de 25 m en 45 m hoog is. Over het hydrologisch functioneren en de nutriënten- en koolstofbalansen van deze bossen is nog niet veel bekend, maar het is duidelijk dat er verschillen zitten in het functioneren van deze bossen.

Dit onderzoek vond plaats in de Cuieiras en Campina natuurparken in de Staat Amazonas, en in een ontbost weidegebied in Rondônia, in Brazilië. De studie is uitgevoerd in zeven verschillende topografische, bodem- en bosgebieden, met name in een ontbost weidegrondgebied, in gedrongen en hoger campina heidewoud op witte zanden in het Campina Reservaat, en langs een topografisch transect met broekbos en *campinarana* overgangsbos op zandgronden in het dal langs de rivier, gevolgd door tropisch regenwoud op de kleiige gronden op de helling en het plateau in het Cuieiras Reservaat.

De doelstellingen van dit promotieonderzoek waren gericht op:

- studie van de bodemrespiratie en het modelleren van de temperatuursafhankelijkheid aan de oppervlakte van de bodemrespiratie voor verschillende bos- en bodemtypes;
- bepaling van de waterbalans, koolstof en nutriëntenafvoer uit een campina heidewoud ter vergelijking met die van het tropisch regenwoud;
- evaluatie van mogelijke klimaatveranderingseffekten (droogte) op de bodem door vergelijking van bodemrespiratie, nutriënt- en koolstofwaarden in een kunstmatig gedraineerd broekbos op zandige grond met die van bodems onder een *campina* heidewoud, en;
- bepaling van de variatie op de lange termijn van de biomassa en bladoppervlakteindex (LAI) van seizoenal weidegrond, en vergelijking van de LAI met die verkregen voor het de verschillende bossen in het regenwoud ecosysteem.

De wetenschappelijke kennis opgedaan in dit onderzoek verschaft nieuwe inzichten in het functioneren van verschillende typen bos (broekbos, campina, *campinarana*, regenwoud) binnen het regenwoud ecosysteem in het Amazonegebied.

9.1 Bodemrespiratie

Hoewel bodemrespiratie een belangrijke rol speelt in de koolstofcyclus van het regenwoud in het Amazonegebied, zijn er slechts weinig metingen aan gedaan. In deze studie hebben we automatische bodemrespiratiemeetsystemen gebruikt om tijdseries van bodemrespiratie en bodemtemperatuur op 5 cm diepte te verkrijgen met een meetinterval van 30 minuten.

Het effect van het insteken van de ring in de bodem werd allereerst bepaald om de meetmethode te verbeteren. Vervolgens is gekeken naar de invloed van neerslag op de bodemrespiratie en de ruimtelijke variaties van CO₂ emissies in centraal Amazone door metingen te verrichten in de zeven verschillende bostypen. Het insteken van de ring in de bodem zorgde voor verhogingen van de CO₂ emissies van 13% en 20%, gedurende een periode van vier tot zeven

uur na het insteken, in respectievelijk zandige en lemige bodems. De bodemrespiratie uit lemige bodem lijkt derhalve wat gevoeliger voor dit mechanische effect te zijn. Dit experiment laat zien dat als de bodemrespiratiemetingen direct na het insteken van de ring gedaan worden, deze kunnen leiden tot een overschatting van wel 20%.

De bodemrespiratie vertoonde geen verandering voor neerslaghoeveelheden onder de 3 mm. Voor hogere neerslaghoeveelheden werden verschillen in respons gemeten tussen de lemige en zandige bodems. Bij neerslaghoeveelheden tussen de 3 en 8 mm werd een stijging van 18% waargenomen tot twee uur na de regenbui, gevolgd door een daling van 15% gedurende de volgende 15 uur. Regenhoeveelheden boven de 8 mm veroorzaakten een onmiddellijke daling van 22–27% in de bodemrespiratie tot 18 uur na de bui. De zandige bodems reageerden veel sneller op neerslag en het effect was veel korter. Voor neerslaghoeveelheden tussen de 3 en 10 mm werd een toename van 34% waargenomen in de respiratie flux gedurende twee uur, waarna de flux weer afnam tot de normale waarde tijdens de volgende twee uur. Neerslaghoeveelheden boven de 15 mm zorgden voor een snelle afname van de flux waarna deze weer terugkeerde naar de oorspronkelijke waarde in de volgende vijf uur.

De bodemrespiratie fluxen verschilden sterk tussen de verschillend bostypen. De laagste waarden ($0,9\text{--}1,2 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) werden gemeten op kale bodem in het gedrongen campina heidewoud op de witte zanden in het Campina Reservaat. Fluxen in het hogere heidewoud in het Campina Reservaat waren hoger tot $3,8 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$. In het Cuieiras Reservaat, waren de gemeten fluxen voor het regenwoud op de kleiige bodems op de helling en het plateau $2,4$ en $3,2 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$, respectievelijk. De bodemrespiratie was hoger in de zandige bodems in het dal onder broekbos ($4,1 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) en waren het hoogste in het campinarana bos ($6,0 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$).

Schattingen voor de bodemrespiratie fluxen van de Cuieiras en Campina natuurparken werden verkregen door de waarden gevonden voor de verschillende bostypen te combineren met de ruimtelijke verspreiding van deze bostypen, zoals verkregen uit Shuttle radar Topographic Mission (SRTM) data. Voor het Cuieiras Reservaat met de broekbossen, campinarana, en het regenwoud op de hellingen en plateaus werd een waarde van $3,8 \pm 0,8 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ berekend. Deze waarde was hoger dan de $3,1 \pm 0,8 \mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ die berekend werd voor de combinatie van kale bodem, kort en hoog heidewoud in het Campina Reservaat. Omdat deze waarden gecorrigeerd zijn voor ruimtelijke verschillen gerelateerd aan verschillen in grondsoort en vegetatie kunnen deze als representatiever worden beschouwd dan die welke zijn gepubliceerd in eerdere studies die niet expliciet rekening hielden met de verschillen in bijdragen van de verschillende soorten bos. Deze studie toont aan dat voor het modelleren van CO_2 respiratie van niet-homogene gebieden informatie over het functioneren van de verschillende landschappelijke eenheden onzekerheden in koolstofemissies kunnen verminderen.

Geautomatiseerde bodemrespiratie metingen geven een beter inzicht in de

mogelijke factoren die bodemrespiratieprocessen beïnvloeden. De belangrijkste factor, behalve in perioden met neerslag, was de bodemtemperatuur, die 90% van de variatie verklaarde in regressieanalyse. Neerslag leek alleen bij te dragen door het veroorzaken van een snelle daling van de bodemtemperatuur tijdens de infiltratie, waarna dit fysische proces de respiratie beïnvloedde.

Bladoppervlakte-indices (LAI) werden gemeten met een Li-Cor 2000 in de verschillende bossen. De LAI varieerde tussen 2,5 en 4,1 in de lage en hoge heidewouden in het Campina Reservaat, respectievelijk. De LAI was hoger voor de bossen in het Cuieiras Reservaat, variërend tussen de 4,9 in het campinarana bos tot 5,5–6,1 voor het regenwoud op de helling en het plateau.

De bodemrespiratiemetingen voor de zeven verschillende bostypen werden ook geanalyseerd op verschillen tussen locaties, en de oorzaken hiervoor, via van een modelmatige benadering. Ensemble gemiddelden van uurlijkse fluxen werden berekend voor natte en droge seizoenen (voor zover deze werden gemeten) van de 30-minuten interval tijdreeksen. Deze waarden werden verwerkt met een analytisch model waarin een schatting van de bodemoppervlaktetemperatuur werd gemaakt uit de waargenomen temperaturen op 5 en 10 cm diepte.

Relaties werden afgeleid tussen de amplitude en fase van de bodemrespiratie R_s en de temperatuur op verschillende dieptes in de grond. In vergelijking met het gebruik van bodemtemperaturen op 5 cm en 10 cm diepte verminderte het gebruik van de door het model berekende oppervlaktetemperatuur de hysteresis tussen de T en R_s sterk en verbeterde de determinatiecoëfficiënt voor de meeste bosgebieden.

Het gebruik van een eenvoudige exponentieel model om de bodemrespiratie uit te drukken als functie van de oppervlaktetemperatuur verbeterde de correlatie tussen de bodemtemperatuur en R_s op dagelijkse tijdschaal. De Q_{10} oppervlaktetemperatuurgevoeligheid varieerde van een lage 1,3–1,7 voor heidewoud in het Campina Reservaat tot een hoge 7,7 voor de verzadigde bodems in het dal van het Cuieiras Reservaat. De verkregen R^2 -waarden varieerden tussen de 0,42 en 0,92. Q_{10} waarden voor het regenwoud op kleigrond op de helling en het plateau waren respectievelijk 3,8 en 5,2. Verschillen in samenstelling van het bos en het bodemtype resulteerden in waargenomen verschillen in temperatuur vertragingen in relatie met CO_2 emissie, met de langste tijd waargenomen voor het traag groeiende heidewoud.

9.2 Bepaling van de waterbalans en de koolstof- en nutriëntenexport uit het Campina Reservaat

Waterbalans studies voor de Amazone zijn bijna altijd gebaseerd op geïntegreerde micro-meteorologische en stroomgebied studies voor tropisch regenwoud, en bieden geen informatie over specifieke vegetatietypen in de regio, zoals voor

broekbossen of campina heidewoud. De koolstofcyclus werd op vergelijkbare wijze via geïntegreerde aanpakken bestudeerd om te beoordelen of het regenwoud ecosysteem in de toekomst netto koolstof blijft opnemen of gaat verliezen. Om informatie te krijgen over het functioneren van een campina heidewoud werd een waterbalans en nutriënten en koolstofexport studie uitgevoerd in een klein stroomgebied in het Campina Reservaat. Deze studie publiceert de eerste schattingen van de waterbalanscomponenten voor een gedrongen en hoger campina heidebos stroomgebied, door combinatie van een waterbalans-, micro-meteorologische en sapstroomstudie, en geeft informatie over de export van opgeloste koolstof en enkele nutriënten.

Gedurende de studieperiode van één jaar (18 maart 2007 tot 19 maart 2008) bedroeg de neerslag 3.054 mm, ongeveer 800 mm boven het langjarige gemiddelde voor deze regio. De verdamping bedroeg 1.532 mm ($4,1 \text{ mm dag}^{-1}$), waarvan 899 mm onder droge omstandigheden plaatsvond. Neerslag interceptie bedroeg 15,6% van de bruto neerslag, uitgaande van een lage stamafvoer fractie gebaseerd op literatuurwaarden voor heidewoud. Transpiratie door het gedrongen heidewoud, gemeten via de sapstroom in deze periode, was 475 mm, wat neerkwam op ongeveer 60% van die van het heidewoud als landschapseenheid. De oppervlakteaflow bedroeg 485 mm, terwijl 1071 mm werd afgevoerd als regionale grondwater uitstroom. In jaren met een gemiddelde neerslag, zal het interceptie verlies lager uitvallen en kunnen bodemvocht tekorten een rol gaan spelen. Onder die omstandigheden kan verwacht worden dat de verdamping van het campina heidewoud lager zal zijn dan dat van het regenwoud wat minder zal lijden onder bodemvocht tekorten vanwege de diepe bodems. Dit wordt ook gesuggereerd door de hogere afvoercoëfficiënt waargenomen voor het campina stroomgebied.

De totale export van opgeloste organische en anorganische koolstof (DOC en DIC) in de oppervlakteaflow bedroeg 15,3 en 3,5 $\text{g m}^{-2} \text{ yr}^{-1}$, respectievelijk. Dit komt overeen met een dagelijkse export van 41,8 mg m^{-2} en 9,5 mg m^{-2} . De afvoeren van DOC en DIC in grondwater waren hoger met 31,7 g m^{-2} en 5,5 g m^{-2} , respectievelijk. De totale hoeveelheid koolstof die in deze periode van een jaar werd afgevoerd bedroeg 55,9 g m^{-2} . Dit is veel hoger dan de jaarlijkse export uit het stroomgebied van het Cuieiras Reservaat (DOC export $< 20 \text{ g m}^{-2}$). Campina heidewouden komen voor in een groot deel van het Amazonegebied, en bij het bepalen van de koolstofbalans van het Amazonegebied zou rekening gehouden moeten worden met deze verschillen in koolstofuitvoer.

9.3 Effekten van klimaatverandering (droogte) op de bodems in de broekbossen in de rivierdalen.

Het op grote schaal kappen van regenwoud om in de behoefte naar landbouwgrond te voorzien kan veranderingen in de lokale en globale water- en kool-

stofcycli veroorzaken. Voor het Amazonegebied zijn toekomstige reducties in neerslag voorspeld in klimaatmodelsimulaties. Veranderingen in de hydrologische cyclus kunnen leiden tot positieve terugkoppelingen in de koolstofcyclus door afsterven van regenwoud (droogte), veranderingen in de CO₂ huishouding van het regenwoud waardoor de balans tussen opname en afgifte van CO₂ verandert van netto opname nu naar netto afgifte in de toekomst, verminderde biomassa-productie en verminderde verdamping van de overblijvende regenwoud. Bossen op zandgronden bedekken ongeveer 6% van het centrale Amazone landschap. In het centrale Amazonegebied zijn de slecht gedraineerde rivierdal bodems van vergelijkbare zandige textuur als die van bodems onder campina heidewoud, maar deze gebieden zijn nu bedekt door broekbossen. Onder de huidige klimatologische omstandigheden ervaren deze broekbossen geen langere periodes van droogte. Echter, een verlaging van het grondwater in deze dalen onder een drogere klimaat zou kunnen leiden tot biogeochemische veranderingen in de bodem waardoor de groeicondities voor het aan de natte bodems aangepaste broekbos zodanig veranderen dat dit leidt tot een verandering in de samenstelling van soorten van het bos.

In deze studie zijn de ruimtelijke en temporele variaties in strooiselafbraak, bodemchemie en grondwaterchemie, en bodemrespiratie bestudeerd in de verschillende soorten bossen die in het centrale Amazonegebied op zandgronden voorkomen (broekbos, campinarana, campina). Ook werd de gevoeligheid voor droogte van de rivierdal bodems onderzocht in een drainage experiment, waarin de grondwaterspiegel kunstmatig verlaagd werd.

Afhankelijk van de grondwaterstand in de zandgronden, werd een grote natuurlijke variatie waargenomen in de koolstof- en stikstofgehalten van de bodem, de uitwisselbare kation concentraties, de bodemwaterchemie en de strooiselafbraaksnelheden. Strooiselafbraak en bodemrespiratie waren zeer laag in de campina heidewouden, waar de capaciteit van de grove zandgrond om koolstof en voedingsstoffen te behouden extreem laag was. De hoogste bodemrespiratie binnen het regenwoud landschap werd waargenomen in de campinarana bossen, die de overgang vormen van broekbos langs de rivier naar regenwoud op de hellingen met kleibodems. Kunstmatige drainage van de broekbosbodem resulteerde in een verlaging van de grondwaterstand met een bescheiden 0,3 m. Dit veroorzaakte geen waarneembare veranderingen in de bodemnutriënten gehalte, de chemische samenstelling van het bodemwater of de strooiselafbraak, binnen de onderzoeksperiode van een jaar. Echter, wel werden aanzienlijke verlagingen gemeten in het koolstofgehalte in de bovenste 5 cm in de bodem (van 1,4% naar 0,1%) en het stikstofgehalte (van 0,05% naar 0,00%). Ook de bodemrespiratie daalde van 3,6 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ bij aanvang van de drainage experiment naar 2,1 $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ acht maanden later. Deze laatste waarde was vergelijkbaar met die waargenomen in het campina heidewoud.

De drainage studie toont aan dat de broekbosbodem zeer gevoelig is voor veranderingen in de grondwaterstand. Door de relatief korte duur van de studie is het onduidelijk of ook de bodem- en waterchemie in de ondergrond in een

later stadium zouden gaan veranderen, en hiermee een transitie naar een ander bostype, zoals naar een droger campina heidewoud, ingezet zou worden met lagere bodemrespiratie. Aan de andere kant zou het broekbos door verandering in plantensoorten en de ontwikkeling van een wortelmat in de tijd kunnen veranderen in het aangrenzende ecotone campinarana bos, wat een hogere bodemrespiratie heeft. Het hangt waarschijnlijk af van de daling van de grondwaterspiegel welk scenario in de toekomst bewaarheid wordt en het blijven monitoren van de drainage en controle plots zou waardevolle gegevens op kunnen leveren. Het grote verschil in bodemrespiratie tussen campina en campinarana bos impliceert dat voor het opstellen van de toekomstige koolstofbalans van het Amazonegebied aanvullende kennis over de veranderingen van bodem en bos in de rivierdalen ten gevolge van een droger klimaat nodig is.

9.4 Biomassa en Leaf Area Index van een weiland in Rondônia

Het Amazonegebied bestaat uit een mozaïek van ongestoord tropisch regenwoud, secundaire bossen, weilanden voor de veeteelt en landbouwgebieden. In Rondônia zijn regenwouden sinds de jaren 50 in de vorige eeuw gekapt voor de aanleg van weilanden voor de veeteelt. In tegenstelling tot het regenwoud vertoont het grasland een seizoenale cyclus van groei, die leidt tot aanzienlijke verschillen in biomassa en bladbedekking (LAI) in de droge en natte tijd. De LAI is een belangrijke parameter voor het modelleren van klimaatverandering en hydrologie, maar er is weinig bekend over de seizoensgebonden variatie in het Amazonegebied. Deze lange-termijn studie (1999–2005), gedaan op basis van destructieve metingen, was de eerste die de variatie in biomassa productie en LAI van weidegebied in het Amazonegebied beschrijft. De begrazingsintensiteit was ten hoogste ongeveer drie dieren per hectare, en de dominante grassoort was *Uruchloa brizantha*.

De biomassa van levend gras was gemiddeld 2905 kg ha^{-1} over de meetperiode van vijf jaar, terwijl de biomassa van het strooisel 2194 kg ha^{-1} bedroeg, wat een totale gemiddelde bovengrondse biomassa van 5195 kg ha^{-1} opleverde. De maximum maandelijkse biomassa steeg tot 8000 kg ha^{-1} bij afwezigheid van begrazing tijdens het natte seizoen van 2004.

De gemiddelde LAI van het weiland was 2,0. Maandelijkse LAI waarden varieerden zowel met de seizoenen, als met de intensiteit van de begrazing. LAI waarden waren meestal rond de 1,2 tijdens het droge seizoen onder begrazing, maar bleef boven 2,0 als er geen begrazing was. Gemiddelde LAI waarden waren slechts iets hoger tijdens het natte seizoen onder begrazing (1,5–2,0), maar stegen tot 3,3 in afwezigheid van begrazing. De maximaal maandelijks gemeten LAI was 4,0. De bijbehorende albedo van het weiland varieerde van ongeveer 0.20 voor LAI waarden beneden 2,0 tot 0,17

voor LAI waarden boven de 3,0. De lange-termijn metingen geven aan dat de biomassa, LAI en albedo van het weiland sterke seizoensgebondenheid vertoont in de afwezigheid van begrazing, die wordt onderdrukt wanneer begrazing plaatsvindt. Op maandelijkse basis kan de LAI van het weiland variëren tussen de 0,5 en 4,1. Deze hogere LAI waarden van weiland zijn vergelijkbaar met die van het heidewoud in het Campina Reservaat (2,5–4,1).

9.5 Conclusies en aanbevelingen

Deze studie heeft aangetoond dat er aanzienlijke verschillen bestaan in het hydrologisch functioneren en de nutriënten- en koolstofcycli van verschillende bostypen die deel uitmaken van het Amazone regenwoud ecosysteem. Het drainage experiment toont aan dat een verlaging van de waterstand in de valleien directe gevolgen heeft voor de bodemrespiratie, en waarschijnlijk ook voor het broekbos als de grondwaterstand permanent wordt verlaagd. Het is nog onzeker of dit zou leiden tot een verandering van broekbos naar campina heidewoud (met lagere bodemrespiratie en waterverbruik) of naar campinarana bos (met hoge bodemrespiratie). Aanvullende kennis over het functioneren van de verschillende soorten bossen, en de effecten van klimaatverandering (droogte) op de toekomstige ruimtelijke verdeling van de verschillende bossen, moet worden verkregen in lange-termijn studies om de gevoeligheid van het ecosysteem voor verandering, en de effecten op de water en koolstof cycli, te kunnen evalueren.

Chapter 10

Acknowledgements

This research was part of the Large-Scale Biosphere–Atmosphere Project (*LBA*) and supported by the Dutch NWO WOTRO W01.65.217.00 project, conducted as a collaboration between the Department of Earth Sciences of the VU University Amsterdam (VUA), the Netherlands, and the *Instituto Nacional de Pesquisas da Amazônia* (INPA) and the Brazilian *Conselho Nacional de Desenvolvimento Científico e Tecnológico* (CNPq). This support allowed me to dedicate several years of my life to writing this PhD thesis.

The first 3 months of this PhD were based out of Manaus, Brazil organizing the future fieldwork in the Cuieiras and Campina Reserve and also the partnership between VUA and INPA. In this period visits to the laboratories were made to find the potential sample analysis and the fieldwork to understand and to fit the project in this landscape of a mosaic of well-drained dissected plateaus, separated by steep slopes and broad, swampy river valleys. Further visits, spanning the length of a year, were made to organize the hydrological and micro-meteorological equipments at VUA and INPA to start the data collection and analysis in the Central Amazonian region. The 2 years of fieldwork were strongly supported by the cooperation with LBA/INPA. Such collaboration was instrumental to the maintenance of the project and data analysis.

It all began with fieldwork in Ji-Paraná city, Rondônia, Brazil, where after three years I published an article with my scientific mentors and advisers, Dr. Maarten Waterloo, Celso von Randow and Antônio Manzi, who always believed in my potential and who helped put me on the science track. After that, I came to Amsterdam to start my PhD where everything was new and understanding the language and the Dutch habits proved tough (many thanks Dr. A. Araújo who help in my adjustment to the Dutch lifestyle). But it was great to be part of this new world of Karnemelk, Stroopwafels, Haring, Gouda cheese, drop, beers, red light district, Vondelpark, all four seasons, and amazing accidents (leaves on the street). Less pleasant were the winters, but it was a good experience biking in the snow or through the beautiful canals of Amsterdam.

Thanks to Dr. Sampurno, Dr. B. Kruijt co-promotor and Dr. A. J. Dolman my promotor who's *Well done!* praises always kept me positive and enthusiastic. I'd also like to thank Dr. F. Luizão and Dr. A. O. Manzi, Manaus LBA leaders, for their invaluable support throughout several stages of this fieldwork.

A special thanks to my co-promotor Maarten Waterloo whom has always been my scientific support, helping me sustain energy for my research and joining me on trips to Manaus to aid with my fieldwork and the building of the meteo tower and hydrological staff gauge in the *Campina* Reserve. Maarten also taught me how to calculate the total carbon from the soil, soil porosity, and aided in my understanding of the hydrological process in the fieldwork. As though this wasn't enough already he also reviewed part of my articles and continually mentored and guided my career over many *rodizios* (barbecues) as well as Japanese restaurant, where I always used these times to find out my scientific track with this busy guy. I must also thank Rita von Randow and Luciana Soler for sharing all their mutual enthusiasm and dare pertaining to the rewarding PhD. battle.

Thanks to A. Meesters, he reviewed one my article inputing a good science on it.

Thanks to M. Groenendijk and her family, they all play important roles in my life. Vrije Universiteit-VU friends H. Beck, D. van der Muelen, Roxana and Berny, Sorana and Seb Iancu, Michiel, Guido, Arnoud, Richard, Gash, Lieselote, amongst others.

In Manaus I am grateful for invaluable field and logistic support, as well as the general support of INPA. I would particularly like to thank: H. B. Xavier, Paulo (doidinho), A. Huxley, J. Souza, Tomé, Galucio, R. Lima and R. Araújo for their invaluable field supply of vehicle and logistic support, as well as E.L. Schloemp and Roberta Souza for the general fieldwork license support with Brazilian government and INPA. I thank M. B. L. de Oliveira, for invaluable field work help and the scientific discussions about K34 tower data. Thanks also to T. Pimentel from Laboratório Temático de Solos e Plantas from INPA.

Here, I want to thank all the technicians who have contributed to the research preparations am very grateful for the sap flow system and all meteo-hydrological equipment from workshop teams of faculty at Earth and Life Sciences-VU University Amsterdam, specially Hans Bakker who worked two months on the Sap Flow sensors, building and constructing them, and also Ron Lootens for his dedication and invaluable support in designing the sap flow heater box and fixing the photosynthesis system. Michel Groen, who has always help me find the necessary meteo and hydrological equipment in the huge University basement.

I also want to thank Marcos Bolson, one of my best friends who also came to work with me in Manaus. I was very lucky when he agreed to come and work on my project as he did my groundwater samples and chromatography analyses, and his expertise made parts of my project much easier.

Thanks to Alexandre Santos (*in memoriam*), who was like a brother and friend during a difficult time in Manaus when he gave me a place to live as

well as his time in the form of stimulating and positive scientific discussions in that melting city (and always followed by a cold beer).

Thanks to the Lima family, especially Tia Sonia (*in memoriam*), Carlos (tche), Fernanda, Michaeli and Fabiane (*in memoriam*).

To all my friends: A. de Almeida, R. Tartari, H. Salazar, Rai(c)a, M. dos Anjos and Pamela, Suanni, Carla, Antonieta and Amarino and Natalie (review part of the English text), whom have been supportive during the tense writing period of my thesis.

And finally but very special thanks to my friends, Erlon, Nei, Paulinho, Wesley and my great family, my mother Neuza A. B. Zanchi, my father Dorival Zanchi and my brother Fernando B. Zanchi, whom always gave me support on this long journey and who believed in my dreams. I apologize to those whose names might have unintentionally been omitted, but who were also important to this huge project.

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List of symbols and abbreviations

The following list gives a short description of the abbreviation, acronyms and chemical symbols used throughout this thesis

Abbreviation/acronyms:

ABRACOS	Anglo–BRazilian Climate Observation Study
AMC	Atmospheric Mesoscale Campaign
$A(z)$	Temperature amplitude (one–half the difference between maximum and minimum temperatures)
<i>ANOVA</i>	One–Way Analysis of Variance
Ca^{2+}	Calcium
<i>C</i>	Soil carbon
<i>C</i>	Control plot
C:N	Soil carbon–nitrogen ratio
CH_4	Methane
Cl^-	Chlorine
CNPq	Conselho Nacional de Desenvolvimento Cientifico e Tecnologico
CO_2	Carbondioxide
<i>CP</i>	Campina Reserve
<i>CPNR</i>	Campinarana
<i>CR10X</i>	Datalogger Type
<i>CU</i>	Cuieiras Reserve
<i>DR</i>	Drainage plot
<i>DOC</i>	Dissolved Organic Carbon
<i>DIC</i>	Dissolved Inorganic Carbon
<i>DOY</i>	Days of year
E_i	Interception loss
Eijkelkamp	Agrisearch Equipment, The Netherlands
ESECAFLORE	Efeitos da Seca da Floresta experiment
FNS	Fazenda Nossa Senhora
<i>GCM</i>	General Global Circulation Models

GPP	Gross Primary Productivity
GPS	Global Positioning System
HAND	Height Above the Nearest Drainage
HBV	Hydrologiska Byrans Vattenbalansavdelning Model
HCl	Hydrochloric acid
HgCl ₂	Mercuric chloride
HNO ₃	Nitric acid
H ₂ O	Water
INPA	Instituto Nacional de Pesquisa da Amazônia
IPCC	Intergovernmental Panel on Climate Change
IRGA	InfraRed Gas Analyzer
K ⁺	Potassium
K _s	Hydraulic conductivity
KOH	Potassium hydroxide
LBA	Large Scale Biosphere–Atmosphere Experiment in Amazonia
LTSP	Laboratorio Tematico de Solos e Plantas
Mg ²⁺	Magnesium
N	Nitrogen
NH ₄ ⁺	Ammonium
NO	Monoxide of nitrogene
Na ⁺	Sodium
N ₂ O	Dioxide of nitrogene
NO ²⁻	Nitrogen dioxide
NO ³⁻	Nitrate
OM	Organic Material
PAR	Photosynthetically Active Radiation
PC1	Campina Piezometer 1
PC3	Campina Piezometer 3
PC4	Campina Piezometer 4
PC5	Campina Piezometer 5
PC6	Campina Piezometer 6
PC7	Campina Piezometer 7
PO ₄ ³⁻	Phosphate
PVC	PolyVinyl Chloride
Q ₁₀	The factor by which soil respiration is multiplied when soil temperature increases by 10°C
RDM	Rolim de Moura municipality
RMSE	Root mean square errors
R ²	Coefficient of determination
r ²	Coefficient of determination
SF	Stemflow
SHF	Stunted Heath Forest
SRTM	Shuttle Radar Topographic Mission
SO ₄ ²⁻	Sulfate

<i>THF</i>	Tall Heath Forest
<i>T_{mean}</i>	Average temperature over the period of measurement
TVEB	Temperature variance–energy balance method
<i>UNIR</i>	Rondonia Federal University, Brazil
<i>VUA</i>	Free University of Amsterdam
<i>ZF2</i>	Cuieiras Reserve

The following list gives a short description of the symbols used throughout this thesis, together with their units.

Symbol:	Description and unit
<i>EC</i>	Electrical Conductivity [$\mu \text{ S.cm}^{-1}$]
<i>ET</i>	Evapotranspiration [mm]
<i>E_d</i>	Actual evapotranspiration under dry canopy conditions [mm]
Efflux	Soil emission [$\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$]
<i>DOC</i>	Dissolved Organic Carbon [mg L^{-1}]
<i>DIC</i>	Dissolved Inorganic Carbon [mg L^{-1}]
<i>H</i>	River water level [m]
<i>L</i>	Leakage [mm]
LAI	Leaf area index [$\text{m}^2 \text{ leaf surface m}^{-2} \text{ ground surface}$]
<i>M_T</i>	Stands for the total dead material [kg ha^{-1}]
<i>T</i>	Soil temperature [$^{\circ}\text{C}$]
<i>TF</i>	Throughfall [mm]
<i>T_s</i>	Surface temperature [$^{\circ}\text{C}$]
ω	Radial frequency [$2*\pi/\text{daylength} (\text{rad s}^{-1})$]
<i>t_{lag}</i>	Time-lag to the model [s]
<i>t_{shift}</i>	Depth-dependent time shift [s]
<i>P</i>	Rainfall [mm]
<i>P_g</i>	Rainfall [mm]
<i>P_G</i>	Gross Rainfall [mm]
<i>P_{FV}</i>	Weight of the leaves from the 20 plants [g]
<i>P_C</i>	Weight of the trunk of the plants [g]
<i>P_{MM}</i>	Weight of the dead material of the 20 plants [g]
<i>Q</i>	Discharge [L s^{-1}]
<i>Q_g</i>	Groundwater discharge [L s^{-1}]
<i>Q_s</i>	Surface runoff [mm]
<i>R₀</i>	Autotrophic respiration [$\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$]
<i>R_s</i>	Soil CO ₂ respiration rate [$\mu\text{mol m}^{-2} \text{ s}^{-1}$]
<i>R_s</i>	Soil respiration [$\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$]
SLA	Specific leaf area [$\text{m}^2 \text{ kg}^{-1}$]
<i>t_{shift,resp}</i>	Time shift for the respiration wave [s]
<i>T_{resp}(t)</i>	Damping depth of soil temperature [$^{\circ}\text{C}$]
<i>Z_{resp}</i>	Depth of optimum soil CO ₂ emission related to the optimum temperature measurement [m]
<i>Z_{5cm}</i>	Depth of 5 cm of the soil temperature measurements [m]
<i>Z_{10cm}</i>	Depth of 10 cm of the soil temperature measurements [m]
<i>T_{5cm}</i>	Average soil temperature at 5 cm [$^{\circ}\text{C}$]
<i>T_{10cm}</i>	Average soil temperature at 10 cm [$^{\circ}\text{C}$]
<i>z</i>	Soil depth [m]
<i>z</i>	Depth of soil respiration dependency [m]