

LONG-TERM MONITORING OF CO₂, H₂O AND ENERGY FLUXES IN
PEATLANDS OF LAKE YENICAGA (BOLU) IN THE CONTEXT OF GLOBAL
CLIMATE CHANGE

by

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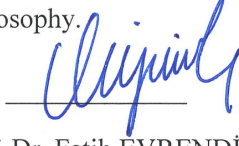
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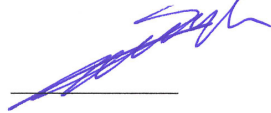
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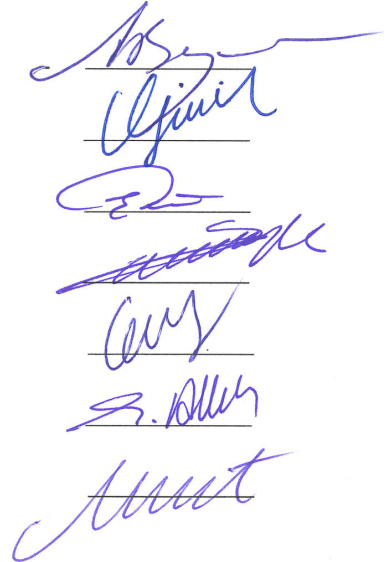
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ABSTRACT

LONG-TERM MONITORING OF CO₂, H₂O AND ENERGY FLUXES IN PEATLANDS OF LAKE YENICAGA (BOLU) IN THE CONTEXT OF GLOBAL CLIMATE CHANGE

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Peatland ecosystems play a crucial role in global carbon (C) cycle as being significant C stocks even though they spatially cover only 5% of the terrestrial biosphere. Human-induced disturbances including global climate change threaten their ability to continue to behave as a C sink. This study of quantification of carbon dioxide (CO₂), water and energy fluxes is the first one conducted for a peatland (Yeniçağa peatland) in Turkey for a monitoring period of three years (from August 2010 to June 2013).

CO₂, energy and water fluxes were measured using an EC technique, in addition to the ancillary measurements of soil and litterfall C:nitrogen (N) ratios, litterfall lignin content, aboveground biomass, and decomposition rate. Mean peat soil C and N contents were 23.3% and 1.7%, respectively. The decomposition constant (-0.006) of soil organic matter (SOM) showed a high decomposition rate with low quality litter (C:N = 18.8, Lignin = 7.2%), and thus, a positive effect on CO₂ efflux. Aboveground biomass peaked in the months of June and August.

Hourly net ecosystem exchange (NEE) varied between -1.0 and 1.5 mg CO₂ m⁻² s⁻¹. The peatland acted as a sink in only June for 2011, a sink from May to July for 2012 and a sink in only May for 2013. The cumulative annual NEE was 352.7, 283.8, 252.4, and 142.6 g C m⁻² for 2010, 2011, 2012, and 2013, respectively. Ecosystem respiration (R_{eco}) and gross primary productivity (GPP) were partitioned using an algorithm for partitioning carbon components. The cumulative annual R_{eco} was always higher than GPP which indicated the high CO₂ source strength of the peatland. The correlation analysis of the partitioned carbon components (NEE, R_{eco} , and GPP) with environmental variables pointed to their highest correlations with air temperature (T_{air}), vapor pressure deficit (VPD) and evapotranspiration (ET) for all the years. As a result, the regression models obtained for GPP and R_{eco} showed higher r^2_{adj} values ($52\% < r^2_{adj} < 78\%$) than those for NEE ($36\% < r^2_{adj} < 61\%$). The water budget of the peatland showed a constant state of water deficit as $ET > PPT$ for the study period. Energy balance closure of the EC flux system was assessed using net radiation (R_n), soil heat flux (G), latent heat flux (LE) and sensible heat flux (H) components and found to be 0.62 and 0.64 for 2011 and 2012.

ÖZET

KÜRESEL İKLİM DEĞİŞİKLİĞİ BAĞLAMINDA YENİÇAĞA GÖLÜ TURBALIK ALANLARINDA (BOLU) UZUN DÖNEMLİ CO₂, H₂O VE ENERJİ AKILARININ İZLENMESİ

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Turbalık alanlar karasal biyosferin sadece%5'ini oluşturmalarına rağmen, önemli karbon stokları olmalarından dolayı küresel karbon (C) döngüsünde önemli bir rol oynamaktadır. Küresel iklim değişikliği de dahil insan kaynaklı bozulmalar, turbalıkların karbon yutağı olma özelliğini sürdürmelerini tehdit etmektedir. Bu çalışma ile Türkiye’de ilk kez bir turbalık alanda (Yeniçağa turbalık alanı) CO₂, su ve enerji akılarının 3 yıl boyunca (Ağustos 2010’dan Haziran 2013’a kadar) izlenmesi gerçekleştirilmiştir.

CO₂, su ve enerji akıları Eddy kovaryans yöntemi ile ölçülmüş buna ek olarak toprak ve döküntü örneklerinin C:N oranı, döküntü lignin içeriği, toprak-üstü biokütle ve çürüme oranı yan ölçümlerle yapılmıştır. Turba toprağında ortalama C ve N içeriği sırasıyla %23,3 ve % 1,7'dir. Toprak Organik Maddesi'nin (TOM) çürüme katsayısı (-0,006) düşük döküntü kalitesiyle (C:N = 18,8, Lignin = %7,2) yüksek çürüme oranı olduğunu göstermiş ve bu durum CO₂ akışına pozitif yönde etkide bulunmuştur. Toprak-üstü biokütlesi Haziran ve Ağustos aylarında zirve değerlerine ulaşmıştır.

Turbalık alanda saatlik net ekosistem CO₂ alışverişi (NECA) -1,0 ve 1,5 mg CO₂ m⁻² s⁻¹ arasında değişmiştir. Turbalık alanın CO₂ yutağı olduğu aylar; Haziran 2011, Mayıs - Haziran 2012 boyunca ve Mayıs 2013 olarak belirlenmiştir. Kümülatif yıllık NECA 2010, 2011, 2012 ve 2013 yılları için sırasıyla 352,7, 283,8, 252,4, ve 142,6 g C m⁻²'dir. Karbon bileşenlerini ayırmak için bir algoritma kullanılarak ekosistem solunumu (ES) ve brüt birincil verimlilik (BBV) belirlenmiştir. Kümülatif yıllık ES her zaman BBV'den yüksek olup, bu durum turbalık alanın yüksek CO₂ kaynağı olduğunu göstermektedir. Karbon bileşenleri ile çevresel değişkenler arasındaki korelasyon analizleri en yüksek korelasyonun hava sıcaklığı (T_h), su buharı basıncı eksikliği (SBBE) ve evapotranspirasyon (ET) değişkenleri arasında olduğunu işaret etmiştir. Sonuç olarak, elde edilen regresyon modelleri BBV ve ES için (%52 < r²_{adj} < %78) NECA'ya göre (%36 < r²_{adj} < %61) daha yüksek r²_{adj} değerleri vermiştir. Turbalık alanın su bütçesi ölçüm döneminde ET > PPT olmasıyla sürekli bir şekilde su eksikliği olduğunu göstermiştir. Eddy kovaryans (EC) akı sisteminin enerji dengesi eşitliği net radyasyon (R_n), toprak ısı akısı (G), gizil ısı akısı ve hissedilebilir ısı akısı bileşenleri ile belirlenmiş ve 2011 ve 2012 için 0,62 ve 0,64 olarak bulunmuştur.

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ABBREVIATIONS LIST

AET	: Actual Evapotranspiration
EC	: Eddy Covariance
ET	: Evapotranspiration
G	: Ground Heat Loss
GHG	: Greenhouse Gases
GPP	: Gross Primary Productivity
H	: Sensible Heat Flux
LE	: Latent Heat Flux
LULC	: Land Use and Land Cover
NEE	: Net Ecosystem exchange
NEP	: Net Ecosystem Production
PET	: Potential Evapotranspiration
PPT	: Precipitation
R_a	: Autotrophic Rspiration
R_{eco}	: Ecosystem Rspiration
R_h	: Heterotrophic Respiration
R_n	: Net Radiation
RF	: Rainfall
SOC	: Soil Organic Carbon
SOM	: Soil Organic Matter
SWC	: Soil Water Content
T_{air}	: Air Temperature
T_s	: Soil Temperature
VPD	: Vapor Pressure Deficit
WB	: Water Budget
WL	: Water Table Depth
β	: Bowen Ratio

CHAPTER I

1. INTRODUCTION

Peatlands have a significant role in the global carbon cycle owing to their ability to sequester carbon (C), thus reducing atmospheric carbon dioxide (CO₂) concentration. Peatlands may act as a sink or source of CO₂ depending on their biogeoclimatic characteristics and human-induced disturbance regimes. Thus, long-term measurements of net ecosystem CO₂ exchange (NEE) between peatlands and the atmosphere are critical to understand sink or source behaviors of peatlands. In the following paragraphs; definition and importance of the peatlands, how peatlands are related to greenhouse gases (GHGs) and climate change, CO₂ cycle in peatlands, and Eddy covariance (EC) measurement technique for CO₂ fluxes are touched on, respectively. Finally, the aim of this study is introduced.

Peatlands are wetland ecosystems where production rate of soil organic matter (SOM) exceeds decomposition and mineralization of SOM (Parish et al., 2008; Donato et al., 2011). Suppressing of SOM decomposition and mineralization is due to high water table that causes anoxic conditions and results in the accumulation of partly decomposed SOM called peat. Peatlands are of great importance in terms of global climate change as they can serve as carbon (C) stocks or sinks. Although

peatlands are of small extent globally (*ca.* 5-8%), they play a significant role in global C cycle. Boreal and subarctic peatlands cover only 3% of the earth's land area, but they store 30% of the world's soil C as peat (Gorham, 1991; Turunen et al., 2002). Owing to the enormous amounts of C stored in the peatlands (may contain up to 450×10^{15} g C, (Roulet, 2000)), great attention has been paid to the role and quantification of C sequestration and fluxes.

Climate is the most important driving variable in determining C sink and source behaviors of peatlands. According to the IPCC (2007) report, CO₂ concentration, the most important anthropogenic greenhouse gas (GHG), has increased globally by about 100 ppm (36%) over the last 250 years. The increased GHG concentrations in the atmosphere are reported to cause global climate change (IPCC, 2007). Climate change exerts quantitatively and qualitatively significant changes on ecosystem structure such as hydrology (Beever et al., 2012), vegetation (Weltzin et al., 2003), and soil (Lal, 2004) and ecosystem function such as biogeochemical cycles (Zepp et al., 2007). Climatic change involves increased air temperature (Hansen et al., 2006), increased atmospheric GHG concentrations (Friedlingstein et al., 2006; Cayan et al., 2007), changes in precipitation regime (Solomon et al., 2007), and more extreme weather events (Diffenbaugh et al., 2005). Thus, climate change is expected to influence peatland C sequestration, and thus, C sink or source strength of peatlands (Belyea et al., 2004; Parish et al., 2008; Strack, 2008). Particularly, in (semi-)arid ecosystems, lowering of water table may quickly change a peatland ecosystem from a C sink to a C source since peat is exposed to aerobic conditions which cause an increase in the rate of SOM decomposition and mineralization (Bubier et al., 2003; Strakova et al., 2012; Grant et al., 2012).

Decomposition and mineralization of SOM by soil microorganisms generate CO₂ under aerobic condition, methane (CH₄) under anaerobic condition, and dissolved organic carbon (DOC) with water contact (Limpens et al., 2008). NEE quantifies difference between CO₂ assimilation by photosynthesis and CO₂ release by total ecosystem (autotrophic and heterotrophic) respiration (R_{eco}) (Lasslop et al., 2010). The direct measurement is possible for NEE, but is not possible for gross primary productivity (GPP) and R_{eco} components using an EC technique. The partitioning of NEE into the GPP and R_{eco} fluxes is important to understand and quantify dynamics of C cycle. Thus, the flux partitioning models have been applied to differentiate GPP and R_{eco} components (Reichstein et al., 2005).

EC method allows for direct and continuous measurements of CO₂, energy and water (H₂O) fluxes between the surface and the atmosphere and was first applied in the 1980s in limited field campaigns. Once the reliability and power issues of the EC instrumentation were tackled in the 1990s, long-term continuous measurements based on EC have been increased across the different biome types of the world (Baldocchi, 2003). Advantages of EC technique mainly stem from being a direct measurement technique, and an area-based rather than point-based measurement which may cover an area of hundred square meters to several square kilometers (Aubinet et al., 2012). Continuous measurements of CO₂, energy and water dynamics of peatlands across the world have been carried out during the growing season (e.g. Shurpali et al., 1995) or for all year long (e.g. Aurela et al., 2002). Many peatlands have been found to be net CO₂ sinks (Nordstroem et al., 2001; Lafleur et al., 2003; Aurela et al., 2004; Lund et al., 2010), also some showed both sink and source properties in different years (Shurpali et al., 1995; 1999; Arneth et al., 2002).

Despite various EC studies about peatlands around the world especially in the subarctic area, there is no EC-based study about CO₂ and H₂O cycles of peatlands in Turkey. The objective of this dissertation is to determine CO₂, energy and H₂O fluxes for Yeniçağa peatland (Turkey) over three years (2010 and 2013) using the EC system. The following research questions are addressed:

- What is the composition of carbon (C), nitrogen (N) and lignin contents of peat soils and plant residues?
- What is rate of changes in plant biomass and decomposition and mineralization of SOM?
- Is Yeniçağa peatland a CO₂ sink or source over the measured time period?
- What are GPP and R_{eco} flux components of NEE? In other words, how much CO₂ is used by GPP and released by R_{eco} by Yeniçağa peatland?
- What are the main environmental controls over NEE, GPP and R_{eco} ?
- Is it possible to model temporal dynamics of NEE and its components as a function of environmental variables?
- What are temporal dynamics of water and energy fluxes over the measurement period?
- Do energy fluxes comply with energy balance closure?

CHAPTER II

2. LITERATURE REVIEW

2.1. Peatlands, greenhouse gases and climate change

Joosten & Clarke (2002) defined peat as “sedentarily accumulated material consisting of at least 30% (dry mass) of dead organic material”, and a peatland as “an area, with or without vegetation, with a naturally accumulated peat layer at the surface”. One of the characteristic properties of peatlands is that the water table remains close to the surface throughout a year which causes anoxic and cool conditions and accumulation of organic detritus. The depth of the accumulated SOM is generally reported to be > 30 or 40 cm or up to several meters (Gorham, 1991) but this depth can vary according to countries’ accepted definitions.

There are two types of peatlands: bogs and fens. The sources of water and mineral for ombrotrophic peatlands (bogs) are precipitation, dust, sea spray, and airborne deposition. Bogs are characterized as mineral deficient peatland type because there is no nutrient supply from groundwater or adjacent ecosystems. However, minerotrophic peatlands (fens) are fed by rainwater as well as surface run-off water. The run-off water from mineral soil provides basic cations (e.g. Ca, Mg, and Na) to

fens and thus fens are also called minerotropic peatlands (Dion and Nautiyal, 2008).

Peatlands cover almost 3% (an estimated area of 4 million km²) of the Earth's land surface (Strack, 2008). Peatland areas for Asia, South America, North America, Australia, Europe, Africa and Turkey are presented in Table 2.1. Peatlands can be source or sink of CO₂ which is the one of the most important greenhouse gases (GHG). Peatlands are also related with other powerful GHG which are methane (CH₄) and nitrous oxide (N₂O). Peatlands may act as a CH₄ source (Lai, 2009) and a N₂O source due to their typical temperature and water table regimes (Maljenen et al., 2010).

Table 2.1. Areas of peatlands (Dion and Nautiyal, 2008)

	Area (10 ⁴ km ²)
Asia	24.4
South America	11.2
North America	174
Australia	2.4
Africa	5.8
Europe	95.7
Turkey	0.024

The natural process of greenhouse effect is illustrated in Figure 2.1. Changes in GHG concentrations in the atmosphere have a significant implication for the Earth's energy budget, and thus, surface temperature. Earth's climate is driven mainly by the sensitive balance between the incoming solar radiation and the outgoing thermal radiation (Bolin, 2008). Roughly one-third of incoming solar (shortwave length) energy is reflected as longwave length energy back into space by the atmosphere (Casper, 2010). The remainder two-thirds is absorbed by the Earth's surface and the atmosphere and emitted back into space more slowly as longwave radiation (Casper,

2010). Most of this thermal radiation is absorbed by the atmosphere and emitted in all directions (Figure 2.1.). The energy absorption increases with a rise in atmospheric GHG concentration (NAST, 2001), thus leading, on average, to a net increase in the earth's surface temperature. This natural process of greenhouse effect is the main reason for the temperature range allowing for the existence of life on the Earth (Ruddiman, 2008). The average temperature of Earth's surface is about 14.6°C (McMichael et al., 2003). Without the greenhouse effect, the average global air temperature would drop from 14.6°C to -18°C; a temperature much too cold to support the diversity of life that exists today on the planet.

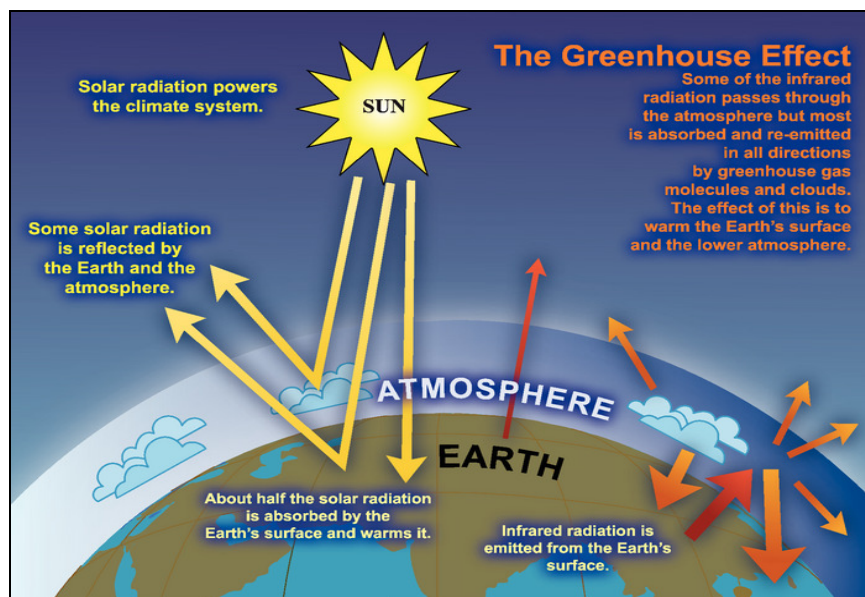


Figure 2.1. An idealized model of the natural greenhouse effect (IPCC Fourth Assessment Report: Climate Change, 2007).

According to 2007 report of Intergovernmental Panel on Climate Change (IPCC), the increased atmospheric GHG concentration (strengthened greenhouse effect) causes global climate change. Global mean surface temperatures have risen by $0.74^{\circ}\text{C} \pm 0.18^{\circ}\text{C}$ over the last 100 years (1906–2005), with the rate of warming over the last

50 years being almost twice the rate over the last 100 years (IPCC, 2007). The stabilization of atmospheric GHG concentrations is critical to ecosystem and human health and well-being. Atmospheric CO₂ concentration has increased globally from a pre-industrial value of about 280 ppm to 395 ppm in 2013 (Figure 2.2.).

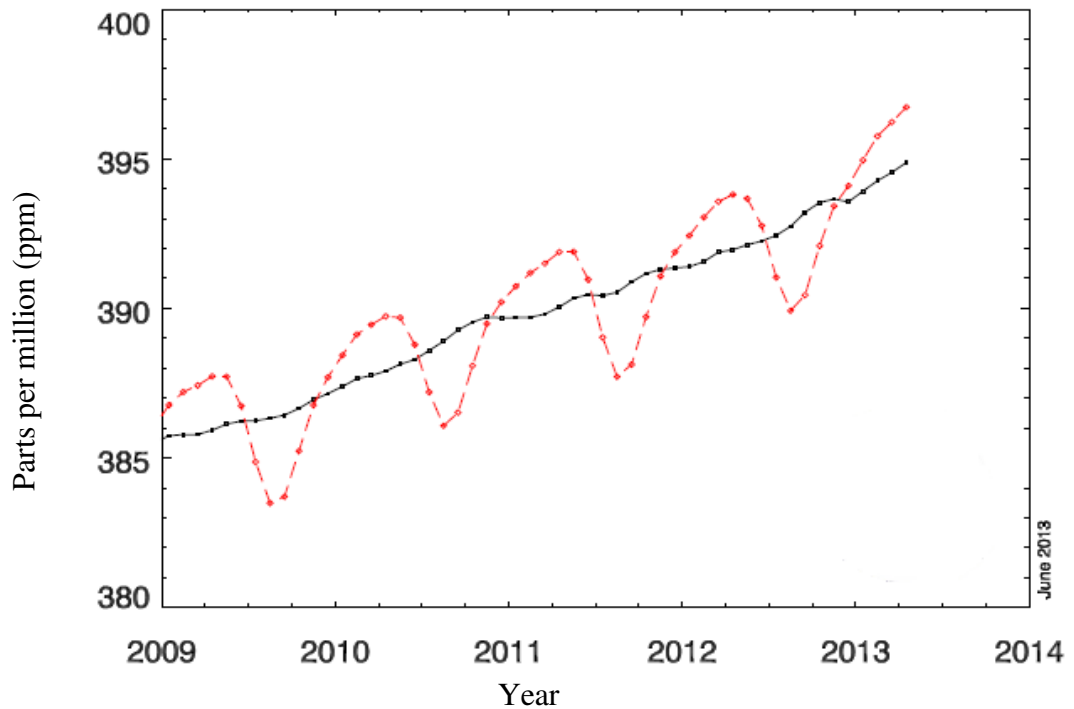


Figure 2.2. Global CO₂ concentrations from 2009 to June of 2013. The global Monitoring Division of NOAA/Earth System Research Laboratory has measured CO₂ for several decades at a globally distributed network of air sampling sites. The dashed red line represents the monthly mean values; the black line represents the same after correction for the average seasonal cycle. (<http://www.esrl.noaa.gov/gmd/ccgg/trends/global.html>).

In this respect, peatlands play a great role in the stabilization of atmospheric CO₂ concentration because they have the potential to serve as CO₂ sinks or sources depending on human-induced disturbance regimes. For example, northern peatlands store between 200 and 450 x 10¹⁵ g C (Gorham, 1991; Turunen *et al.*, 2002),

equivalent to approximately 10-30% of global soil C stocks and to 30 and 75% of the pre-industrial C pool stored in the atmosphere (Strack, 2008). The global area of the peatlands was reported to have reduced significantly by at least 10 to 20% (Joosten & Clarke, 2002). Therefore, research on peatland ecosystems is urgently needed to understand and quantify their role in global C cycle, and thus, to tailor preventive and mitigative measures towards sustainable management of peatlands, and consequently, the stabilization of the atmospheric GHG concentration.

Effects of human-induced disturbances including global climate change on CO₂ metabolism of peatlands are of great importance since CO₂ release from peatlands to the atmosphere could act as a positive climatic feedback (Strack et al., 2006). That is, increased atmospheric CO₂ concentration leads to warming, thus increasing decomposition and mineralization of soil organic carbon (SOC) pools which in turn increase the CO₂ release to the atmosphere. However, considering effect of change in one single climatic factor such as increased temperature or CO₂ fertilization due to climate change on one single component of an ecosystem such as SOC storage or plant biomass is most likely to lead to misunderstanding of interactions and dynamics that climate change would alter for peatlands at the ecosystem level.

Interactive effects of increased nighttime versus daytime temperature, CO₂ fertilization, changed precipitation regime, and increased extreme weather events must be monitored and quantified on both ecosystem structure and function so that a more realistic picture of ecosystem and human responses can be captured in order to mitigate climate change. However, impacts of climate change on peatlands are mostly focused on changes in water cycle because the principal factor that controls peatland ecosystems is the hydrological regime (Gorham, 1991; Roulet et al., 1992;

Moore, 2002). When water table level decreases (in growing season), peat soil is subjected to aerobic conditions that accelerate the decomposition and mineralization rates of SOC which shifts the ecosystem status from C sink to C source (Bubier et al., 2003). In addition, increasing soil temperature causes an increase in CO₂ emissions in the peat soils (Jun-Qin et al., 2009; Wang et al., 2010). For northern peat soils, related studies showed that CO₂ emissions increased in response to climate change (Moore, 2002).

2.2. Carbon cycle in peatland ecosystems

Carbon budget of a peatland is closely associated with the following components: CO₂ and CH₄ exchanges between the biosphere and the atmosphere, and dissolved organic carbon (DOC) losses in streams (Moore et al., 1998). CO₂ exchanges are focused in this study as the dominant part of C cycle presented in Figure 2.3.

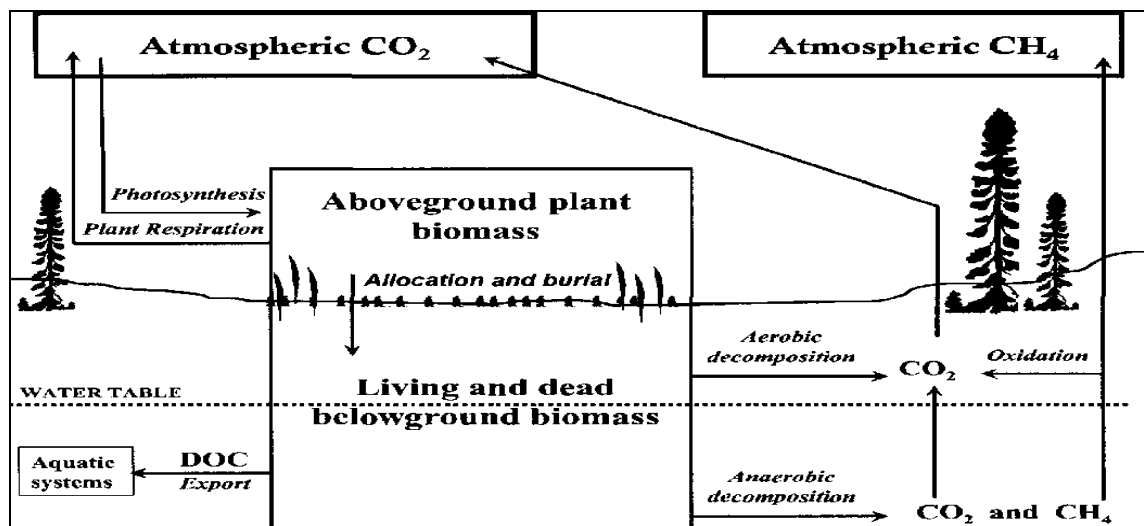


Figure 2.3. The peatland carbon cycle. Vegetation fixes atmospheric CO₂ through primary production and a portion of this carbon is transferred to belowground biomass (peat). Aerobic and anaerobic decomposition transforms solid and dissolved organic carbon into CO₂ and CH₄ that diffuse to the atmosphere (Conlin, 2008).

The first step in the cycle is the plant uptake of atmospheric CO_2 by photosynthesis. Photosynthetic rate is controlled by the interaction of biotic and abiotic factors such as photosynthetically active radiation (PAR), evapotranspiration, precipitation, groundwater level (WT), vegetation, atmospheric CO_2 concentration, air temperature, and human-induced and natural disturbance regime. Between 49% and 70% of CO_2 is released to the atmosphere by autotrophic respiration (Lorenz et al., 2010). Plants (autotrophs) respire in order to maintain the complex tissue components, to take up ions and to construct new cells (Waring and Running, 2007). Soil respiration (heterotrophic respiration) is the result of an aerobic decomposition and mineralization of SOC by which dead plant residues (whether be belowground or aboveground) are converted from coarse SOC to fine SOC and then are released as inorganic CO_2 to the atmosphere by microorganisms, respectively (Figure 2.4) (Luo et al., 2006).

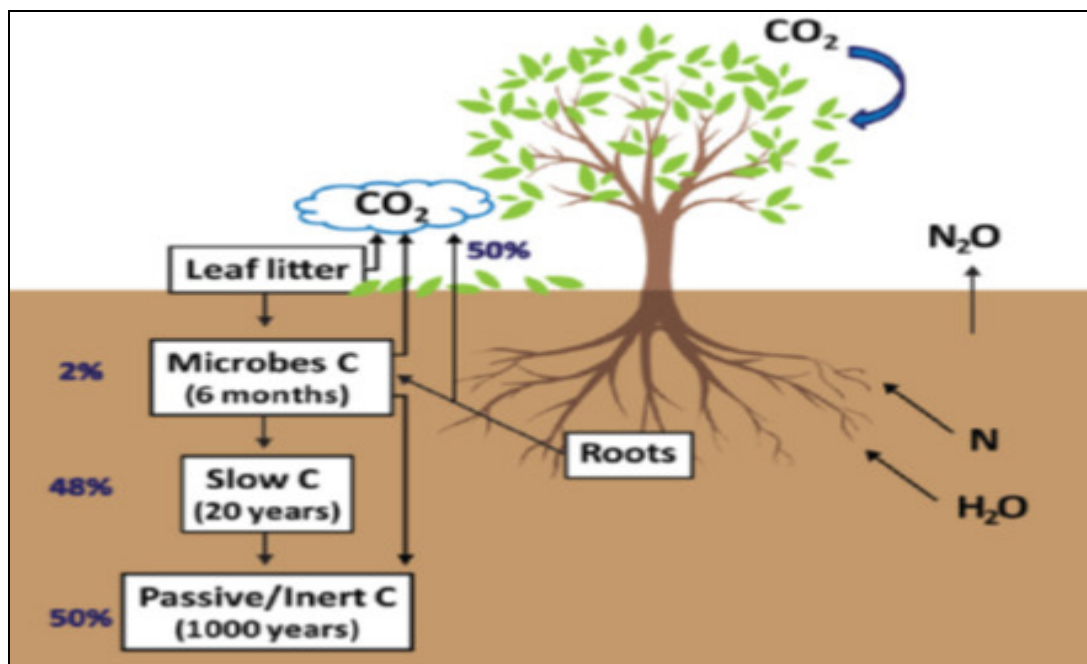


Figure 2.4. Conceptual pools and fluxes often used in process-oriented SOM models (Stockmann et al., 2013).

In peatlands, water level divides the peat profile into aerobic and anaerobic sections. Compared to other ecosystems, decomposition and mineralization of SOC is slow and dependent upon oxygen availability, microbial activity, temperature, and vegetation type (Ise et al., 2008). For example, C dynamics is affected by vegetation type in that plant biological productivity, litter decomposability, and association with fungi are the main vegetation-related driving variables (White II, 2008). These variables affect not only photosynthetic and respiration rates but also C amount available for mobilization as DOC from the peatland (Koehler, 2010). Also, air and soil temperature plays a key role in controlling the rates of biogeochemical processes in peat soils (Smith et al., 2003; Wisser et al., 2011; Rowson et al., 2013). Plant uptake of CO₂, and soil respiration all have a temperature range in which microbial and plant enzyme activities operate at optimum (Fenner et al., 2005; Wang et al., 2010). The temperature dependence of soil respiration is defined by the Q_{10} equation which states that rate (R) of a reaction increases for every 10 °C rise in the temperature (T) (Davidson et al., 2006). This equation can be used for the comparison of different rates of microbial processes and their response to an increase in temperature. Some studies showed that soil respiration rate in peat increased by 1.5 - 2 times for a 10°C increase in temperature (Dorrepaal et al., 2009; Aguilos et al., 2011).

Above the water level, above and belowground litter is decomposed and mineralized under aerobic conditions, which produces the majority of CO₂ emissions. Anaerobic decomposition is carried out by microorganisms and methanogenicarchaeae, and results in CH₄ emissions. Methanogenesis in peatland ecosystems occurs universally below the water table where conditions are anoxic (Kettunen, 2002; Wania, 2007).

Carbon cycle can be briefly summarized as consisting of CO₂ uptake by photosynthesis and CO₂ emissions by soil and plant respirations. Gross primary production (GPP) of an ecosystem represents total CO₂ uptake by photosynthesis without the consideration of autotrophic respiratory CO₂ losses (Ciais et al., 2010). Total ecosystem respiration (R_{eco}) is the sum of both autotrophic (R_a) and heterotrophic (R_h) respiration (Figure 2.5.).

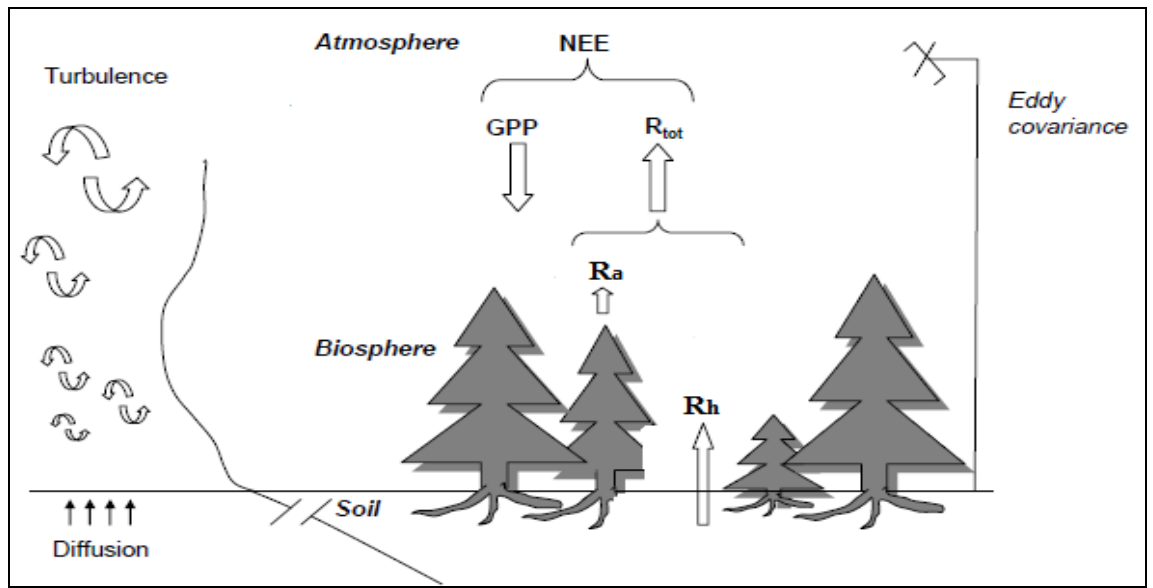


Figure 2.5. Schematic view of the transport mechanisms of gases (on the left hand side), CO₂ flux components; net ecosystem exchange (NEE), gross primary productivity (GPP) and total ecosystem respiration (R_{eco}), and an eddy covariance tower on the right hand side used in this study for flux measurement. R_a : autotrophic respiration. R_h : heterotrophic respiration (Lohila, 2008).

Net ecosystem production (NEP) was originally defined by Woodwell and Whittaker (1968) as the difference between the amount of organic C gained by photosynthesis (GPP) and the amount of organic C lost by total ecosystem respiration (R_{eco}).

$$NEP = GPP - R_{eco} \quad (2.1)$$

$$R_{eco} = R_a + R_h \quad (2.2)$$

Measurement of NEP is challenging due to different temporal scales of CO₂ intake or uptake processes involved including respiration and photosynthesis. For example, C can be released from an ecosystem in the form of volatile organic compound (VOC) on a short-term timescale. On a long-term timescale, part of the annual NEP may be lost from the ecosystem as dissolved or particulate organic or inorganic C to the groundwater, as CH₄ to the atmosphere (Smith et al., 2010) or carried away by the wind via exports of C in pollen and seeds (Eugster et al., 2008). Therefore, NEP estimates have high uncertainties (Chapin et al., 2006) although some of these components are of minor importance depending on ecosystem type (Ciais et al., 2010). Thus, another way to quantify NEP is to measure vertical net flux of CO₂ above canopy level which gives net ecosystem exchange (NEE). Net ecosystem exchange is equal to NEP plus inorganic C flux.

$$-NEE = NEP + \text{inorganic C flux} \quad (2.3)$$

The minus sign for NEE denotes C uptake by the biosphere. Inorganic C loss or gain can be in the form of leaching into groundwater and the conversion of CO₂ (Barr et al., 2004). NEE is the net CO₂ exchange between the biosphere and the atmosphere, that is, vertical CO₂ flux from the ecosystem to the atmosphere (Baldocchi, 2003; Urban et al., 2007). In a conventional metrological notation, net fluxes of CO₂ into the atmosphere are considered positive (+), while net uptake by the ecosystem is considered negative (-). With the negative NEE sign, the equation is adopted in the following form;

$$NEE = -GPP + R_{eco} \quad (2.4)$$

If C gain exceeds C loss, then the photosynthesis process dominates, and the ecosystem becomes a CO₂ sink. If the opposite occurs, then the respiratory processes dominate, and the ecosystem becomes a CO₂ source. During the daytime hours, measured NEE is the sum of photosynthesis and autotrophic and heterotrophic respiration. In the nighttime, NEE represents ecosystem respiration because photosynthesis ceases.

The processes can be briefly described as:

Daytime:

$$NEE = R_a + R_h - GPP = R_{eco} - GPP$$

Night:

$$NEE = R_a + R_h = R_{eco}$$

Consequently, negative NEE values denote the removal of CO₂ from the atmosphere by the ecosystem, whereas positive NEE values indicate the reverse.

2.3. Water cycle in peatland ecosystems

Water and energy dynamics of a peatland ecosystem control CO₂ exchange. For peatlands, water is of major importance since peat formation is closely coupled to water-saturated conditions. Necessary and sufficient conditions are required for peat formation to occur where they occur actually. One of the necessary conditions to that effect is that precipitation or groundwater influx should be higher than evapotranspiration (Martini et al., 2006). Water is one of the important factors for the classification of peatlands. As mentioned in Section 2.1., bogs and fens types have

different properties due to their water and nutrient sources. Traditionally, the water budget for the peatlands can be defined by using Eq. (2.5):

$$\text{PPT} + \text{G} = \Delta(\text{M}, \text{R}) + \text{Q} + \text{ET} \quad (2.5)$$

where PPT: precipitation, G: groundwater inflow, M: soil water storage, R: surface storage, Q: runoff and ET: evapotranspiration

Generally for bogs, the water budget (WB) is defined with subtraction of water losses through stream discharge (SD) and ET from water input through PPT:

$$\text{WB} = \text{PPT} - \text{ET} - \text{SD} \quad (2.6)$$

Evapotranspiration is a major component of the water budget of both natural and managed bogs and fens (Drexler et al., 2004). Evapotranspiration term is used to describe the total process of water transfer from vegetated land and bare soil surfaces to the atmosphere (Chandran et al., 2006). ET from wetlands is controlled by three main factors: the amount of energy used for the process, the ability of the atmosphere to hold and transport water vapour, and the ability of vegetation and soil to transport water through the atmosphere (Scottocornola, 2007).

Potential ET (PET) is the maximum evapotranspiration rate that can occur where water availability is not limiting. PET is dependent upon the amount of solar radiation, temperature, humidity and wind speed, and effects of soil and plant are ignored. Actual ET (AET) is the rate of water loss on a normal day and is limited by water availability, plant physiology and atmospheric conditions (Hughes et al., 2001).

Penman defines PET (mm day^{-1}) as evaporation from an open water surface, wet vegetation or bare soil (Penman, 1948) (Eq.2.7). According to Penman formula PET calculations require such meteorological variables as net radiation, air temperature, air water vapor deficit, and wind speed. The calculation of PET using Penman formula is relatively easy because the required variables and parameters can be easily measured under field conditions.

$$PET = \frac{mR_n + \rho_a c_p (\delta_e) g_a}{\lambda_v(m + \gamma)} \quad (2.7)$$

where m : slope of the saturation vapor pressure curve (Pa K^{-1}), R_n : net irradiance (W m^{-2}), ρ_a : density of air (kg m^{-3}), c_p : heat capacity of air ($\text{J kg}^{-1} \text{K}^{-1}$), g_a : momentum surface aerodynamic conductance (m s^{-1}), δ_e : vapor pressure deficit (Pa), λ_v : latent heat of vaporization (J kg^{-1}), γ : psychrometric constant (Pa K^{-1})

Penman-Monteith method for PET (Monteith, 1965) refers to the use of an Eq. (2.8) for computing water evaporation from vegetated surfaces (e.g. Hughes et al., 2001; Chandran et al., 2006).

$$PET = \frac{\Delta(R_n - G) + \rho c_p \frac{D}{r_a}}{\Delta + \gamma(1 + \frac{r_s}{r_a})} \quad (2.8)$$

where PET: mm day^{-1} , G : soil heat flux ($\text{MJ m}^{-2} \text{day}^{-1}$), ρ : atmospheric density (kg m^{-3}), c_p : specific heat of moist air ($\text{kJ kg}^{-1} \text{°C}^{-1}$), r_a : aerodynamic resistance (s m^{-1}), r_s : surface resistance (s m^{-1})

The Priestley-Taylor method (PT) (Priestley and Taylor, 1972) is a simplification of the Penman equation used to allow calculations of ET which assumes that an air mass moving over a homogeneous, well-watered surface would become saturated. Under these ideal conditions, ET would reach a state of equilibrium (Priestley and Taylor, 1972). The PT model is based on the assumption that the effect of turbulence is small compared to the effect of radiation. The method is appropriate for use when detailed meteorological measurements are not available. This model has been used for calculation of ET in wetlands (e.g. Petrone et al., 2001; Parmentier et al., 2009; Clark et al., 2010).

The most common methods for measuring ET in wetlands are the Bowen ratio energy balance (e.g. Kellner, 2001; Abtew, 2004; Peacock and Hess, 2004) and Eddy covariance (EC) (e.g. Eugster, 1997; Twine et al., 2000; Petrone et al., 2001; Kurbatova et al., 2002; Acreman et al., 2003).

Bowen ratio energy balance method calculates ET from the ratio of sensible heat (H) to latent heat (LE) (known as Bowen ratio) derived from the ratio between atmospheric temperature and humidity gradients measured a few meters above vegetation (Abtew et al., 2013). Bowen ratio can be used as an indicator to define the regional climate. For example, if H is larger than LE , then $\beta > 1$ and more turbulent energy is returned to the atmosphere as sensible heat. This situation usually indicates a dry climate. Conversely, when $LE > H$ then $\beta < 1$, which shows a more humid climate (Lafleur, 2008).

Bowen ratio is based on the following energy balance Eq. (2.9) (Burba et al., 1999):

$$R_n - G = H + LE \quad (2.9)$$

where R_n : net heat gain from radiation (W m^{-2}), G : ground heat loss (W m^{-2}),
 H : sensible heat loss (W m^{-2}), LE : latent heat loss (W m^{-2})

Latent heat can be calculated from a rearrangement of Eq (2.9):

$$LE = \frac{R_n - G}{\beta + 1} \quad (2.10)$$

where the Bowen Ratio (β):

$$\beta = \frac{H}{LE} \quad (2.11)$$

within a few meters of the surface H and LE may be expressed as follows:

$$LE = \frac{-\rho C_p}{\gamma} k_v \frac{\Delta e}{\Delta z} \quad (2.12)$$

$$H = -\rho C_p k_H \frac{\Delta T}{\Delta z} \quad (2.13)$$

where Δe : change in vapour pressure with height (kPa), ρ : atmospheric density (kg m^{-3}), C_p : specific heat of air ($\text{kJ kg}^{-1} \text{K}^{-1}$), ΔT : change in temperature with height (K), z : height (m), γ : psychrometric constant (Pa K^{-1}), k_H and k_v : turbulent transfer coefficients for heat and vapour ($\text{m}^2 \text{s}^{-1}$)

The Bowen ratio is calculated by combining Eq. (2.12) and (2.13). Generally k_H and k_v are assumed to be equal, and thus, they cancel one another in Eq. (2.14):

$$\beta = \gamma \frac{\Delta T}{\Delta e} \quad (2.14)$$

Therefore Bowen ratio can be calculated from Eq (2.14) measuring the change in temperature and vapour pressure which is used in Eq. (2.10) to achieve latent heat or evapotranspiration.

2.4. Eddy Covariance Methodology

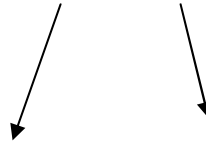
2.4.1. Eddy Covariance Technique

Eddy covariance (EC) is a micrometeorological technique that provides a direct measurement of CO₂ and H₂O fluxes between the vegetation canopy and the atmosphere layer. EC allows for a continuous estimation of NEE for a long-term (Su et al., 2008; Tang et al., 2012; Mauder et al., 2013). A flux is the transfer of a quantity per unit area per unit time (Delgado et al., 2012). An eddy flux is the transfer of a quantity by the turbulent motion of air (Delgado et al., 2012).

The technique is based on atmospheric turbulence generated from the movement of air immediately above the earth's surface. The result of the atmospheric turbulence is the formation of the eddies which cause transportation of the heat and mass between the surface layer and free atmosphere. In the atmosphere, upward and downward turbulent motions of the air occur and EC technique determines the net difference of the air mass moving across the canopy-atmosphere interface. The basic principle of the EC technique is that the vertical flux of a material in the turbulent surface layers is proportional to the covariance of the vertical eddy velocity and the material's concentration (Baldocchi, 2003).

The measurement of the flux with using this technique is fulfilled by statistical analysis of the instantaneous vertical mass flux density ($F = w\rho_c$, $\mu\text{mol m}^{-2} \text{s}^{-1}$), using Reynolds' rules of averaging (Reynolds, 1895). According to the Reynold's rule, the instantaneous mass flux can be expressed as the sum of the mean term of trace gas concentration varying over long periods and the turbulent parts varying over short periods of tens of seconds. For example, CO_2 concentration in the vertical direction can be written as follows:

$$F_c = \overline{w\rho_c} + \overline{w'\rho_c'}$$



where; mean vertical flux eddy flux

w: wind speed in vertical direction (m s^{-1}); ρ_c : CO_2 density ($\mu\text{mol m}^{-3}$)

As indicated before, EC measurements provide us with vertical turbulent flux at the measurement point. Theoretically, the flux measured by EC at the measurement height is assumed to equal NEE between the ecosystem and the atmosphere. Some assumptions and corrections are available behind the EC measurements. The surface of interest is assumed to be a horizontally homogeneous and flat terrain. The horizontal flux divergence is much smaller than vertical flux divergence and is assumed to be neglected (Finnigan, 1999). The mean vertical wind speed at or near the surface (especially above short vegetation) is small and assumed to be neglected. There is assumed to be no mass movement of dry air in the vertical direction over a suitable interval of time which means that $\overline{w} = 0$ and so the equation becomes:

$$F_c = \overline{w'\rho_c'}$$

Then, the turbulent flux (F_c) is often defined as NEE (Aubinet et al., 2012). The trace gas density (ρ_c) can vary due to the variation of the air density caused by temperature and pressure gradient difference along the vertical direction. Webb correction (Webb and Leuning, 1980) was defined to minimize the effect of the fluxes on air density (Liebethal and Foken, 2003).

In a similar way, LE and H fluxes can be calculated with the EC technique as indicated in the following formulas:

$$LE = L_v \overline{w' \rho_v'}$$

where L_v : latent heat of vaporization (kJ kg^{-1}), w' : vertical wind velocity fluctuations (m s^{-1}), ρ_v' : water vapour density fluctuation (g m^{-3})

$$H \cong \rho_a c_p \overline{w' T'}$$

where ρ_a : moist air density (kg m^{-3}), c_p : specific heat capacity of moist air ($\text{J kg}^{-1} \text{°C}^{-1}$), T' : temperature fluctuations (°C)

2.4.2. Advantages and limitations of EC technique

Under ideal conditions, EC technique supplies accurate and reliable flux measurements; however, in practice, EC method has some limitations. One limitation is the underestimation of night fluxes by the EC method. During the calm nights, the turbulence is not strong enough because the surface layer becomes stable. This weakness prevents CO_2 transport from being measured. This leads to CO_2 accumulation on the surface layer and underestimation of NEE measurements

between the ecosystem and the atmosphere (Aubinet et al., 2012a). As a result, the night measurements fail and data gaps occur. The solution for the nighttime problem is to apply an appropriate gap-filling procedure such as interpolations for short-term data gaps (Baldocchi, 2003), empirical models for longer gaps built as a function of explanatory variable(s) (Falge et al., 2001).

Another weakness of the EC method is the difficulty to determine footprint of a flux tower. The footprint was defined as the ground area under the control of EC measurements (Schmid, 2002). The EC system should measure the turbulent fluxes that represent an ecosystem; in other words, the field of view should be determined properly. The exact location of a sensor is not a problem for a homogeneous surface because the fluxes come from all directions of the surface area equally. However, the situation is not same if the surface is not homogeneous. In a heterogeneous landscape, the flux may change with wind direction, atmospheric stability, and surface roughness (Marcolla et al., 2005). Three different models have been developed to determine footprint area for a flux tower: Eulerian analytical models, Lagrangian-stochastic dispersion models, and Large eddy simulation model (Scottocornola, 2007). In a homogenous topography, the height of the tower can be used to determine footprint using the ratio of 1:100 (1 vertical meter per 100 horizontal meters) as a rough rule of thumb (Campbell Scientific, 1998). Additionally, several post-processing operations and corrections are needed when calculating the EC flux from the high-frequency data collected (Aurela, 2005). Although the EC technique has some limitations, it has many advantages that can be briefly expressed as follows (Baldocchi et al., 1998; Baldocchi, 2003):

- providing long-term, direct and continuous measurements of CO₂ and H₂O

fluxes

- allowing for measurements over large sampling areas instead of point measurements
- not disturbing ecosystem conditions

2.4.3. Data processing and filtering

The Eddy Covariance (EC) method is one of the most direct ways to measure and calculate turbulent fluxes at high frequency (e.g. 10 Hz, 20 Hz) within the atmospheric boundary layer (Aubinet et al., 2012). However, the error and uncertainties associated with the calculation and measurements of the flux are still an outstanding issue (Papale et al., 2006; Mauder et al., 2007). The raw data processing and filtering are important for the quality of the data (Aubinet et al., 2012).

There are several ways to process the high frequency raw data collected by the EC system. The datalogger used in EC system can already apply online processing on the raw data or some software tools can be used for that purpose (e.g. Eddypro, EdiRe, EddySoft). The software tools are utilizable for both processing and filtering of the raw data, which can be called as post-processing. The following post-processing steps can be applied by using the software tools: despiking of raw data, correction axis rotation (Lee et al., 2004; Wilczak et al., 2001), frequency response corrections (Moore, 1986), sonic-temperature sensible heat flux corrections for humidity (Schotanus et al., 1983) and the density corrections on CO₂ and H₂O measurements (Webb et al., 1980). The online processing with datalogger programs does not execute any filtering steps on the raw data, only it performs the followings:

statistics for turbulence and meteorological variables (include fluxes, means and standard deviations), wind direction and wind speed, second moments (cross-products) hydrometer oxygen correction, and Webb et al. (1980) correction (Campbell Scientific, 2004-2006), that can be categorized as pre-processing.

Although various attempts have been made to standardize processing methodology (Aubinet et al., 2000; Lee et al., 2004) there is still an on-going discussion on standardization (Lee et al. 2004; Mauder and Foken 2006). The available software programs, EdiRe (e.g. Baum et al., 2008; Thomas et al., 2011), EddyPro (e.g. Ward et al., 2012; Rains, 2013), EddySoft (Schindler et al., 2006; Merbold et al., 2011) have been preferred in the literature to provide the post-processing of the raw data. The difference usages in software tools may cause significant discrepancies in the flux results. The comparisons of the improved software tools for eddy covariance flux resulted in good agreement but not yet a perfect one (5-10% difference for 30-min CO₂ flux values) (Mauder et al., 2007).

The filtering step is applied for the data quality after the data-processing step. Even the raw data are post-processed by the software tools some undesired data may come out. High frequency raw data often contain spikes, dropouts, constant values, and noise due to the instrumental problems (Aubinet et al., 2012). Not only the post-processing process is unstandardized but also the filtering of the processed data has not been standardized. However, the importance of the standardization procedure has been emphasized and the standardization procedure has been improved on (e.g. Papale et al., 2006).

In order to make comparisons of the results more reliable, the strategy of the filtering should be followed in parallel. The strategy was based on two main steps: threshold determination and spike removal. Firstly, threshold values should be determined for the physical parameters (wind velocity range, temperature range, and gas concentrations, respectively) (Aubinet et al., 2012). The main physical problem is the nighttime turbulence (Aubinet et al., 2012a). The eddy covariance technique often underestimates nighttime CO₂ flux under stably stratified atmospheric conditions (eg. Valentini et al., 2000; Gu et al., 2005; Zhilin et al., 2006). Typically the stable atmospheric conditions tend to suppress turbulence at night causing a decrease in atmospheric mixing. In these conditions, the trace gas can accumulate close to the ground, not reaching the sensor height, causing night time gas flux underestimations (Baldocchi, 2003). The practical solution used in the related literature is to filter the nighttime measurements due to the atmospheric turbulence conditions using friction velocity (u^*) as an indicator (Aubinet et al., 2012). The filtering procedure that can be applied for u^* selection was described by Goulden et al. (1996). The flux data are discarded below the defined threshold value for u^* . This threshold value is determined by plotting nighttime flux to u^* data to see the relations. By looking the relations, when there is not linear relationship between the nighttime flux data and u^* , the lower u^* value is defined as a threshold value. Secondly, the standard deviation of the average interval should be checked. The following definition was proposed for the spike determination (Aubinet et al., 2012);

$$\left| X_i - \overline{X_j} \right| \geq D^* \sigma_j \rightarrow spike \quad (2.15)$$

where; X_i =value at a time, X_j = mean value of taken time window, D = discrimination factor (e.g. $D=3.5$), σ_j = standard deviation from the mean value.

The processed data can be filtered from the spikes using Eq.(2.15). The EC fluxes are ready to be analyzed after the filtering step. The algorithm used in this study was based on information mentioned in this section and is introduced in Section 4.2.1.

2.4.4. Gap-filling

Gaps in data coverage occur during the long-term EC measurements because of mainly equipment failure, system maintenance shutdowns, power-failures, unfavorable meteorological conditions and also filtering procedure (Hui et al., 2004; Reichstein et al., 2005; Chen et al., 2012). The EC system provides half hourly or hourly data for 24 h and 365 days. However, 20 to 60% of flux observations are generally reported as missing or rejected (Falge et al., 2001; Hui et al., 2004; Moffat et al., 2007). Therefore, gapfiling should be performed for providing complete datasets.

Falge et al. (2001) detailed gap-filling strategies for EC systems. The available methods can be classified as follows: look-up tables (Falge et al., 2001), a method based on the mean diurnal variations of previous periods (Falge et al., 2001), (non)-linear regression (Aubinet et al., 2000; Rupper et al., 2006), multiple imputation (MI) (Hui et al., 2004; Dragomir et al., 2012), process-based models (Xing et al., 2008) and artificial neural network (ANN) (Archibald et al., 2009; Jarvi et al., 2012; Evrendilek et al., 2013).

In *look-up tables* (LUT) method, the missing fluxes are estimated based on the environmental drivers (such as light and temperature) associated with missing data

(Falge et al., 2001). In practice, a multidimensional table is created where the missing value can be “*looked up*” based on the values of the meteorological drivers. The tables are created according to fixed periods over a year with corresponding fixed intervals for the variables (e.g. six bimonthly periods or four seasonal periods) (Falge et al., 2001). The average flux is calculated for each of fixed periods and the missing data are replaced by the average flux from the related periods.

Mean diurnal variation (MDV) is another gap-filling method based on the temporal auto-correlation of the fluxes (Falge et al., 2001). In this technique, the missing flux at a certain level such as 1-h is replaced by the average flux calculated from the adjacent days (7-14 days) at exactly that time of day (Falge et al., 2001).

Nonlinear Regressions (NLRs) methods rely on the functional dependence on meteorological variables such as temperature and light (Moffat et al., 2007). In this method, a regression equation based on a cause-effect relationship such as between flux and temperature or between flux and light is fitted into a valid dataset which is in turn used to fill the missing fluxes. Various versions for nonlinear regression equations have been proposed (Barr et al., 2004; Desai et al., 2005; Richardson et al., 2006; Qun et al., 2013); however, two different equations are used generally: one for nighttime data and one for daytime data. Nighttime data and daytime data are often estimated as function of temperature and light, respectively (Aubinet et al., 2012). The rectangular hyperbola function in which the driver is light (Eq. 4.3) is used commonly for estimating daytime fluxes (Foke, 2008). For nighttime flux, the most common equation used is Lloyd and Tylor equation (Section 4.2) in which the driver is temperature (Moffat et al., 2007).

Multiple imputation (MI) is a statistical method based on the multivariate correlation. In this method, the missing data are replaced with a set of plausible values instead of replacing with a single value for each missing datum (Hui et al., 2004). The MI method relies on the following three main steps: imputation (creating multiply imputed dataset), analysis (analyzing complete dataset using a standard procedure), and pooling (combining complete data results for inference) (Buuren, 2012).

In the *process-based models*, the flux prediction is based not only on driving variables such as meteorological variables but also on state variables such as soil type, soil texture, soil depth, and canopy height (Moffat et al., 2007). This method application requires knowledge of complex interrelationships of the variables involved in estimating the fluxes for a long-time period (Xing et al., 2008; Aubinet et al., 2012).

The *Artificial Neural Networks* (ANNs) are purely empirical nonlinear regression models (Aubinet et al., 2012). ANN consists of nodes that are arranged in successive layers where the information flows unidirectionally, from the input layer to output layer, through the hidden layers (Hanrahan, 2011). The network is trained by sets of meteorological drivers and associated output flux data (Moffat et al., 2007). After training, the relations of flux with meteorological inputs are mapped onto the weights and the ANN is then used to predict the missing flux values (Moffat et al., 2007).

The comparisons of the gap-filling methods for C fluxes performed by Moffat et al. (2007) showed that all the methods yielded, on average, good results but with a slightly higher performance of ANN. In addition, a gap length of lower than 10 days

resulted in low uncertainty depending on the selection of methods. Thus, it can be concluded that the selection of the gap-filling methods depends on such considerations. For example, MDV method can be used since any meteorological data are not available or LUT is utilizable if some meteorological data are available, or the nonlinear methods can be used since it is validated for the daytime and nighttime data, or advanced statistical methods (e.g. ANN, MI) can be improved if ancillary measurements (meteorological drivers, vegetation type, soil type etc.) are sufficiently available.

2.5. CO₂ fluxes for peatlands in related literature

NEE measurements have been dominantly performed in the arctic (Laine et al., 2006; Laurila et al., 2010), boreal (Lloyd et al., 2010; Strack et al., 2013) and tropic (Hirano et al., 2007; Saunders et al., 2007) regions. In this study, the prevailing climate regime was cool temperate, and therefore, studies in a similar region were compiled in this section for a comparison of annual NEE results. To the author's best knowledge, annual NEE measurements for temperate peatlands were rare and were summarized in a decreasing chronological order in the following paragraphs.

Strilesky et al. (2012) performed a study to measure NEE at a cool temperate peatland of Mer Bleue open bog (45° 24'N, 75° 31'W) for the period of 1 November 2006 to 31 October 2007 and reported that net CO₂ uptake was -104 g C m⁻² yr⁻¹ over a single year. The open bog temperate peatland was a source of CO₂ from October to the end of March and a sink for CO₂ from April to the end of September.

For another temperate peatland (Fajemyr), Lund et al. (2007; 2012) measured CO₂ exchange in southern Sweden (56° 15 N, 13° 33 E) in the long-term period (2005-2009) and reported that the peatland acted as a small sink (-21.4 g C m⁻² yr⁻¹) between August of 2005 and July of 2006 (Lund et al., 2007). CO₂ flux measurements continued to make inter-annual comparisons between 2006 and 2009 showed that the annual NEE balance was 14.3, -29.4, 23.6 and -28.9 g C m⁻² yr⁻¹ for 2006, 2007, 2008 and 2009, respectively (Lund et al., 2012). The inter-annual variation in NEE was large due to differences in NEE during the growing season (late May to early September). Drought for the growing season and higher water table level (WL) in 2006 and 2008 caused the peatland to act as an annual source of CO₂. The sink or source strength was reported to depend on the metrological conditions that affect soil moisture content, temperature and water table depth.

Koehler et al. (2011) stated that an Atlantic blanket bog near Glencar in a temperate region of southwest Ireland (51°55'N, 9°55'W) where mean annual air temperature (MAT: 10.6°C) was slightly higher than those of Mer Bleue (6.7°C) and Fajemyr (6.2°C) peatlands was a net sink of CO₂ for the months of May to September and a net source of CO₂ between October and April in the study period of 1 January 2003 to 31 December 2008. The annual cumulative NEE indicated net CO₂ uptake during the study period and showed a high variation ranging from a minimum of -12.5 g C m⁻² yr⁻¹ in 2006 to a maximum of -84.0 g C m⁻² yr⁻¹ in 2005. The driest summer of 2006 led to the minimum uptake of CO₂, and also a lower CO₂ uptake (-13.5 g C m⁻² yr⁻¹) was obtained in wet and cool summer of 2007 which suggested that the sink strength decreased at extreme weather conditions (a dry or wet summer).

Zhou et al. (2009) pointed out that NEE of a warm temperate wetland at Panjin, China (121° 54'E, 41° 08'N) during 2005 (MAT: 8.6°C) was a net uptake of -65 g C m⁻² yr⁻¹. Nilson et al. (2008) performed NEE measurements between 2004 and 2005 for Degero Stormyr peatland (64° 11'N, 19° 33'E) in the cold temperate region of the country of Vasterbotten, Northern Sweden (MAT: 1.2°C) which led to a net uptake of -55 and -48 g C m⁻² yr⁻¹, respectively. It was reported that there was no extreme weather event and the sites acted as a sink during growing season (May-September) and a source for the rest.

Roulet et al. (2007) revealed that the annual cumulative NEE ranged a minimum of -2 g C m⁻² yr⁻¹ (2000-2001) to a maximum of -112 g C m⁻² yr⁻¹ (2003-2001), with a mean annual cumulative NEE of -40.2 g C m⁻² yr⁻¹ between 1998 and 2004 for a cool temperate Mer Bleue peatland located 10 km east of Ottawa, Canada (45.41°N, 75.48°W) (MAT: 6.0°C). The NEE variation was high during the six-year period, and the period during which sink strength was low coincided with the period during which the water table level was quite high.

Nieveen et al. (1998) measured the CO₂ fluxes from June 1994 to October 1995 for a disturbed temperate peat bog in the Fochteloer area in the north part of the Netherlands. During June, July and August, the bog peat behaved as a sink of CO₂ attributed to environmental conditions such as soil water and vegetation status. However the system behaved as a source from November 1994 to November 1995 at a rate of 97 g C m⁻² yr⁻¹ most probably due to drainage.

Shurpali et al. (1995) showed that a temperate peatland in Minnesota acted as a sink (-32 g C m⁻² yr⁻¹) during a cool and wet summer of 1992 which gave rise to high

photosynthesis and low soil respiration rate. However, the peatland acted as a source ($71 \text{ g C m}^{-2} \text{ yr}^{-1}$) for a warmer and drier summer of 1991. Warmer and drier weather conditions led to a decrease in water table level and to an increase in soil aeration and soil respiration.

In light of the above results reported in the related literature, CO_2 exchange shifted from a source to a sink from late April to May and conversely from September to October. In response to such extreme meteorological events as droughts, flooding, decreased water table level, and drainage, the peatland behavior changed from a sink to a source.

CHAPTER III

3. STUDY SITE & METHODOLOGY

3.1. Study site description

The Yenicaga peatland is located about 38 km east of the city of Bolu (40°47'N', 32°1'E) in the western Black Sea region of Turkey (Figure 3.1). The climate in the Yenicaga region is classified as cool temperate, with a mean annual temperature and precipitation of 10.2°C and 538 mm, respectively (Dengiz et al., 2009). The Yenicaga peatland is reported to contain Devonian and cretaceous limestone, basaltic tuff, lava, and olistolites, with the uppermost layer consisting of tertiary and quaternary formations (Dengiz et al., 2009).

The natural vegetation types of the Yenicaga peatland consist of the following dominant communities: (1) *Phragmites australis* and *Typha domingensis* (41.2 ha); and (2) *Ranunculus lingua*, *Acorus calamus*, *Najas marina*, *Pedicularis palustris*, *Senecio paludosus* (107.8 ha) (Sumer, 2002).

An EC flux tower site was installed about 1 km north of Lake Yenicaga (18 km²) at the elevation of 988 m above sea level on July 12, 2010. The mean vegetation height

around the flux tower is about 0.5 m, and the terrain observed around the flux tower exhibits flat and uniform grasslands.

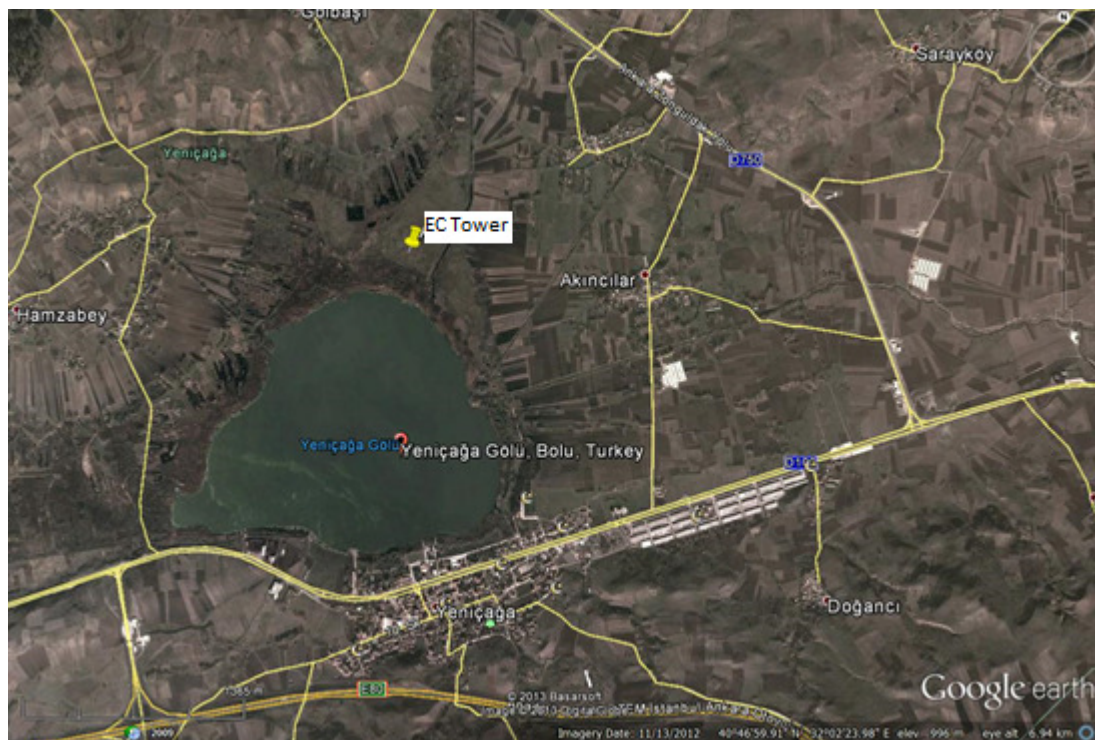


Figure 3.1. Location of the study site “Yenicaga peatland” in northwestern Turkey. EC Tower; Location: 32° 1'44.81"E-longitude, 40°47'24.26"N-latitude, Elevation: 995 m.

3.2. Field sampling procedure

The field sampling was performed to observe the decomposition rate, biomass, productivity, Carbon(C)/Nitrogen (N) and Lignin contents of soil and plant residue around flux towers.

Determination of decomposition rate

The litterbag technique was used for determination of decomposition rate (Moore et al., 2007). In this technique, fresh leaf litter is enclosed in mesh litterbags, placed in

the ground to a soil depth of 5 cm, and collected at periodic intervals to measure mass remaining (Karberg, 2008). Differences between the mass at the beginning of the experiment and that at the different time periods were calculated to evaluate litter decomposition rate (Singh and Gupta, 1977). By incubating the leaves *in situ*, they are exposed to the normal fluctuations in temperature and moisture. The mesh litterbags allow smaller insects as well as microorganisms to access to the leaves. 1 mm mesh size is sufficiently small to prevent losses of litter due to breakage, but sufficiently small to permit the access of decomposers (Sundarapandian and Swamy, 1999). The mass loss of plant residue as a function of time is a first order reaction and the mathematical expression can be written as follows (Barlocher, 2005);

$$\ln\left(\frac{M_0}{M_t}\right) = k * t \quad (3.1)$$

where; M_0 = mass of litter at time 0, M_t = mass of litter at time t, t = time of incubation, k = decomposition rate constant.

In this study, five points around the flux tower were selected and a meshed litterbag of 19 x 25 cm was used at each point. During a year from June of 2012 to June of 2013 every 60 days the samples were collected to measure the mass loss. The experimental details are illustrated in Figure 3.2. Firstly, seven mesh litterbags (mesh size ~1 mm) were prepared for each sampling point. The litter was collected homogenously from each labeled point and transported to the laboratory. At the laboratory, 10 g of seven-subsamples were weighed and six of them were filled into the mesh litterbags. Six mesh litterbags buried under approximately 5 cm below the peat and each mesh litterbags were taken off to weigh at defined time intervals.

Collection dates of the litterbags are presented in Table 3.1. At laboratory, one of the seven-subsamples (10 g) for each point was oven dried at 60°C for 48 hours (Zhou et al., 2009). Next, dried subsamples were weighed and recorded. The conversion factor was used to convert weight of the subsamples into dry weight. The dry weight of the subsamples divided by weight (10 g) gives the conversion factor to be used for the determination of dry weight of other subsamples.

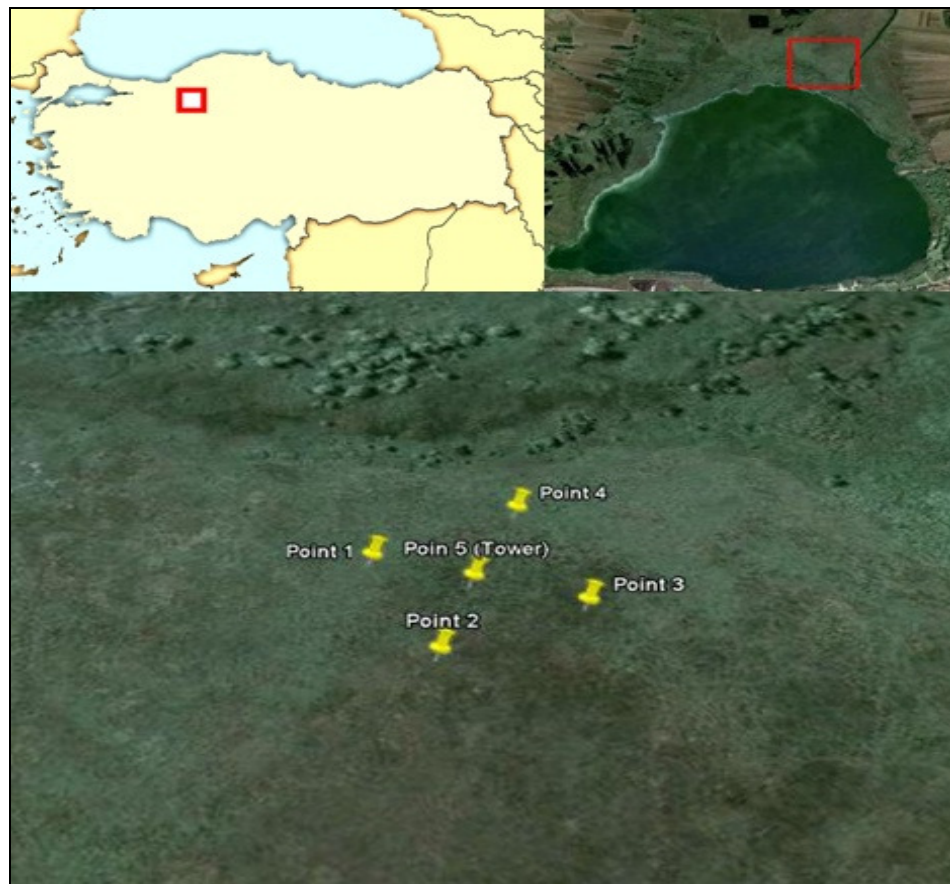


Figure 3.2. Five sampling points are selected around flux tower and six meshed litterbags are buried at each point. A total of 30 meshed litterbags are buried and weight losses are measured during a year at a 60-day interval.

At the end of the experiment, the dry weights of the samples were recorded using the conversion factor for each time interval. The average dry weight was calculated using five points for each time interval. Then, the plot of log (mean dry weight

subsamples) as a function of time gives mass loss decay. The decomposition constant, k , can be calculated using Eq. (3.1).

Aboveground biomass measurement

In situ measurements were made to observe amount of aboveground biomass and rate of plant productivity. This method entails harvesting of total biomass at a specific area and then drying them to estimate dry weight biomass. Monthly aboveground biomass was collected from the five points around the flux tower. Aboveground biomass was harvested over an area of 0.5 x 0.5 m. The sampling for aboveground biomass was carried on nine sampling dates: June 17, August 15, September 15, October 15, November 15, December 15 in 2012, and 21 February, 21 March, 03 May in 2013. Aboveground biomass samples were collected from each point and transported to the laboratory of the Department of Environmental Engineering of A.I.B.U. The samples were oven dried at 60°C for 48 h until a constant dry weight was observed (Zhou et al., 2009). For each sampling date, the same procedure was applied. Results were expressed as g C per area (m²). The amount of C in dry weight aboveground biomass as with dry weight litterfall was estimated using a coefficient of 0.5 to convert dry weight into C content (Evrendilek et al., 2006). Knowledge of aboveground biomass density is useful in determining the amount of standing carbon stocks in ecosystems. Plant productivity refers to the rate of change in plant aboveground biomass per a given area and time. Plant productivity dynamics are important to quantify relationship among net primary productivity, net ecosystem productivity and soil respiration (Trettin et al., 2005).

Analysis of carbon and nitrogen contents

Nineteen peat soil samples to a depth of 0 to 30 cm and litterfall samples were

randomly taken around the flux tower for once. Then, the samples were dried at 60°C for 48 hours, grinded and weighed (1.5-2.0 mg) into tin capsules in order to get them ready for C and N analyses. Carbon and nitrogen contents of the soil and litterfall samples were analyzed using an elemental CHN analyzer (Euro EA3000, EuroVector, Milano, Italy) which was run at 980°C and used Helium as a carrier gas. The calibration was performed with using BCR 71 CRM standard that contains 10.86% N and 40.55% C.

Analysis of lignin content

Lignin content of nineteen litterfall samples collected randomly around the tower was determined using a standard protocol of EN ISO 13906:2008. The following steps were performed to obtain lignin content of litterfall.

1. Prepare Acid Detergent solution (ADS) for dissolving the acid-labile carbohydrates, leaving a fibrous residue that is primarily cellulose and lignin.
ADS: Weigh 49.04 g concentrated H₂SO₄, reagent grade, into a 1000 mL volumetric flask containing 400 mL H₂O. Make up to volume with distilled water. Add 20 g CTAB (Cetyl trimethylammonium bromide), technical grade, and dissolve.
2. Prepare 72% H₂SO₄. Cellulose and other organic matter are solubilized by 72% H₂SO₄ and thus, the remaining is defined as Acid Detergent Lignin (ADL).
3. Capsule tray with 6 positions was used. Dry capsules with lids in an oven at 105°C for at least 30 min, desiccate and cool for at least 5 min.
4. Weigh capsule + lid (W₁), tare and then weigh 1.0 g of grinded litterfall samples to an accuracy of ±0.1 mg (W₂) into each capsule. A blank was run

through the entire procedure to measure any contribution from reagents and residue.

5. Close the lids of the capsules and insert into the capsule tray; place in the acetone solution (120 mL) and agitate for 1-2 min, and repeat three times. Remove tray holder and allow capsules to drain. This step was for a pre-extraction to remove the fat.
6. Place extraction beaker with 350 mL ADS on the hot plate. Place condenser on the top of the extraction beaker and open cold water tap for the reflux of the system. Boil for 60 minutes.
7. At the end of the boiling time, empty the extraction beaker and capsules from the liquid. Fill the extraction beaker with 350 mL boiling water and wash the capsules. Repeat the washing procedure for five times.
8. Dry the capsules in an oven for at least 3 h at $120 \pm 2^\circ\text{C}$.
9. Cool the capsules to room temperature (~30 minutes) in a desiccator and weigh with a precision of ± 0.1 mg (W_3).
10. The capsules are inserted into the capsule tray and put into the beaker with 120 ml 72% H_2SO_4 . Leaved to soak for 3 h.
11. Dry the capsules to room temperature (~30 minutes) in a desiccator and weigh with a precision of ± 0.1 mg (W_4).
12. Weigh the pre-dried ashing crucibles (W_5). Ash the capsules in the ashing crucible for at least 4 h at $600 \pm 10^\circ\text{C}$.
13. Cool the ashing crucibles slowly. Weigh with a precision of ± 0.1 mg (W_6). Also weigh the blank capsule (d).
14. The capsule can lose a small amount of weight during reaction with the reagent. Therefore, correction factor was used to compensate for this loss.

Correction factor (c) was calculated dividing W_4 to W_1 . And the mean of c was estimated as 0.999 as indicated in the standard method.

$$15. \quad \%ADL = 100 \times \frac{W_4 - (W_1 \times c) - (W_6 - W_5 - d)}{W_2}$$



Figure 3.3. A picture of Lignin analysis content of litterfall. Capsule tray with 6 positions in extraction beaker with a condenser was on the hot plate.

3.3. EC and ancillary measurements

Eddy covariance (EC) system consisting of an open-path $\text{CO}_2/\text{H}_2\text{O}$ gas analyzer (LI-7500, Licor Inc., Lincoln, NB, USA) (Figure 3.3.), a 3-D sonic anemometer/thermometer (CSAT3, Campbell Scientific Inc., Logan, UT, USA) (Figure 3.4.), a data logger (CR3000, Campbell Scientific Inc.) (Figure 3.5), and a 3-m tower on which EC flux sensors were mounted. The distance between the LI-7500 and CSAT3 sensors was 0.15 m, with CSAT3 oriented towards the prevailing wind direction (an azimuth angle of 30° from true north) and LI-7500 vertically rotated 15° towards the footprint. Eddy fluxes and associated signals are recorded at 10 Hz, block averaged over one hour (h) and corrected for the effects of fluctuations in air density on $\text{CO}_2/\text{H}_2\text{O}$ fluxes through the online flux computation. EC data were

collected swapping two 2-GB Compact Flash cards at about 15 day intervals. The net radiation (R_n), downwelling and upwelling longwave (4 to 50 μm) and short wave radiation (0.2 to 4 μm) (R_{l_dn} , R_{l_up} , R_{s_dn} , and R_{s_up} , W m^{-2}) were measured using Kipp & Zonen CNR-4 net radiometers (Kipp & Zonen USA Inc., Bohemia, New York) (Figure 3.6). The picture of EC tower in Yeniçağa peatland was shown in Figure. 3.7.

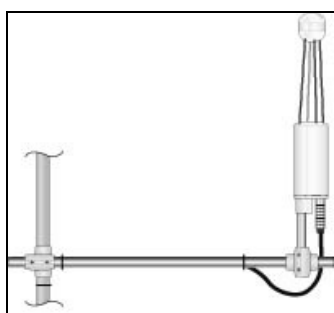


Figure 3.4. LI-7500 Open path $\text{CO}_2/\text{H}_2\text{O}$ gas analyser



Figure 3.5. CSAT 3 three dimensional sonic anemometer



Figure 3.6. CR300 data logger



Figure 3.7. Kipp & Zonen CNR-4 net radiometers (Kipp & Zonen USA Inc., Bohemia, NY, USA)



Figure 3.8. EC tower in Yeniçağa peatland

Groundwater level was measured monthly using a perforated pipe of 1.5 m in length buried near EC flux tower. Air temperature (T_{air} , °C), and relative humidity (RH, %) were measured using a HMP45C probe (Vaisala, Finland). Rainfall (RF, mm), evapotranspiration (ET, mm), soil water content (SWC, %), and mean, maximum and minimum soil temperature (ST , ST_{max} , and ST_{min} , °C) were measured on hourly basis using an ET107 weather monitoring station (Campbell Scientific Inc.). The summarized information for measurement techniques is given in Table 3.1.

Table 3.1. The methodological summary

Variable	Units	Measurement Technique	Measurement duration
CO ₂ flux	mg CO ₂ m ⁻² s ⁻¹	Eddy Covariance (EC), Signals recorded at 10 Hz by datalogger (CR300), and that data logger gives hourly averaged data	Compact flash cards used to collect data at ~15 day intervals
Latent Heat flux (LE)	W m ⁻²		
Sensible Heat flux (H)	W m ⁻²		
Precipitation (P)	mm h ⁻¹	ET107 weather monitoring station. Data recorded hourly	
Evapotranspiration (ET)	mm h ⁻¹		
Soil water content (SWC)	%		
Soil temperature	°C		
Air Temperature	°C	HMP45C probe	
Relative humidity			
R _{l_dn} , R _{l_up} , R _{s_dn} , and R _{s_up} ,	W m ⁻²	Kipp & Zonen CNR-4 net radiometers (Kipp & Zonen USA Inc., Bohemia, NY, USA)	
C and N contents of peat and litterfal	%	Elemental CHN analyzer Euro EA3000, EuroVector, Italy	19 soil and litterfall samples collected randomly around the tower for once
Lignin content	%	EN ISO 13906:2008 standard protocol	
Mass loss (decomposition)	g C m ⁻² d ⁻¹	Litterbag	60 days interval during a year (2012-2013). Initial time= 24/06/2012
Productivity	g C m ⁻² d ⁻¹	Harvesting at five points around the EC tower	June of 2012 to May of 2013, monthly
Water table depth	cm	Buried perforated pipe	January of 2011 to July of 2013, monthly

CHAPTER IV

4. RESULTS

4.1. Meteorological measurements

The rainfall (RF), soil water content (SWC), soil temperature (T_s) and air temperature (T_{air}) were recorded hourly from 01 August 2010 to 19 June 2013. The mean monthly values for each were calculated to constitute Figures 4.1 to 4.5. The mean monthly rainfall and cumulative rainfall from August of 2010 to June of 2013 were presented in Figures 4.1 and 4.2, respectively. The winter months (December to March) showed a similar trend during which mean rainfall was between 0.01 and 0.06 m month⁻¹. As it is expected; the most drought season occurred in the summer (June to September). An exception was that the maximum mean rainfall was recorded in June of 2011. Also, Yeniçağa peatland was flooded from April of 2011 to the mid-June of 2011 and also in April of 2012. The cumulative rainfall was 152.1 mm in 2010, 408.7 mm in 2011, 106.2 mm in 2012, and 150.6 mm in 2013 (Figure 4.2). The annual cumulative comparisons could be made only for the complete annual datasets of 2011 and 2012 which the most arid year was 2012.

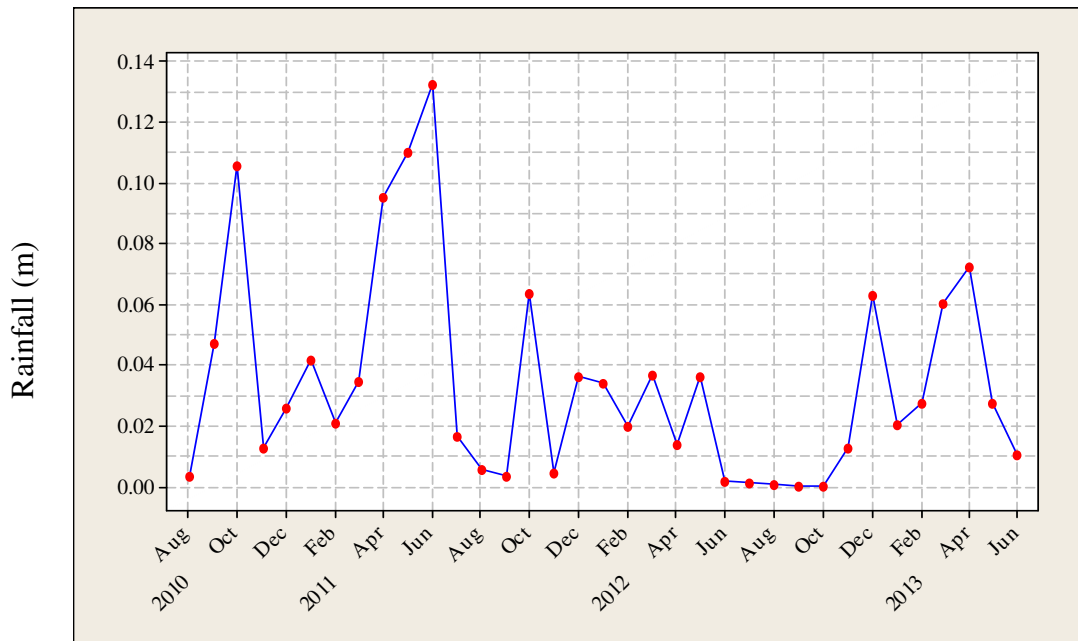


Figure 4.1. Changes in mean monthly rainfall from August of 2010 to June of 2013.

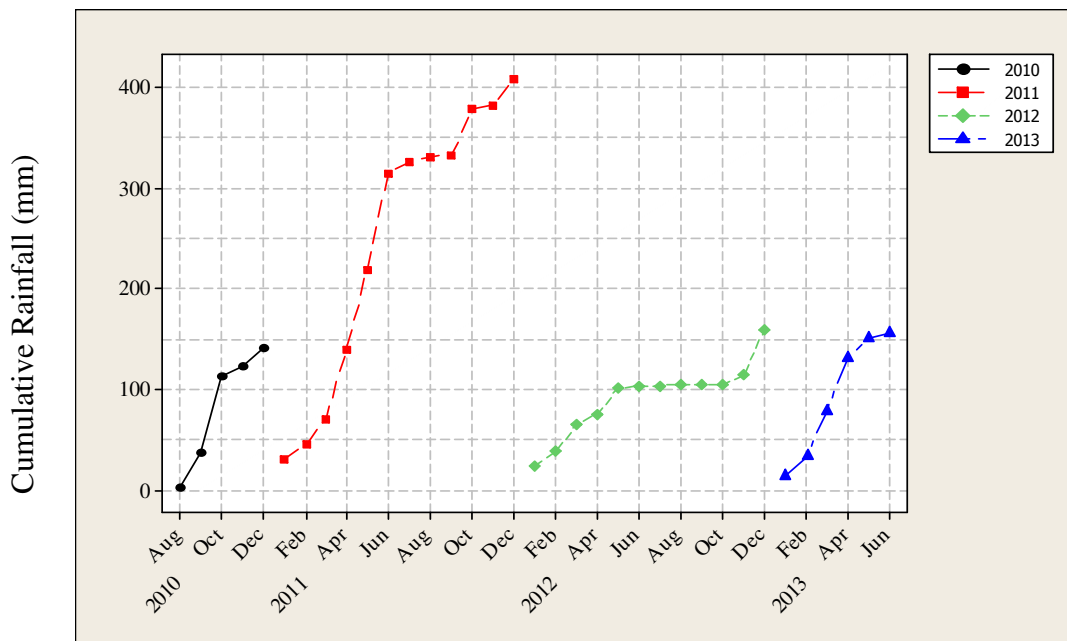


Figure 4.2. Changes in cumulative rainfall between 2010 and 2013.

Soil water content (SWC) showed a sharp decrease from August of 2010 (81.1%) to September of 2010 (69.8%) and then fluctuated between 72% and 74% until the end of March of 2011 (Figure 4.3). The effect of flooding was noticed on SWC in April

of 2011, which increased to 85% and stayed over 85% throughout the flooding period (May to June of 2011). From June of 2011 to June of 2012, SWC did not decrease below 85% but showed a small decrease from June of 2011 to September of 2011 that may be caused by a decrease in RF during that time (Figure 4.1). RF increased in October (Figure 4.1) which was most likely to lead to a slight increase in SWC in November. The relationship between RF and SWC for the period of June of 2012 to October of 2012 showed that the lack of RF decreased SWC sharply up to 63%. Then a little increase occurred in RF during November and December of 2012 (Figure 4.1) but that was not enough to make an increment in SWC. It can be concluded that a sharp increase in RF resulted in an increase in SWC at the subsequent month (e.g. April to June of 2011), and a decrease in RF (e.g. July to October of 2012) caused a decrease in SWC at the subsequent months (e.g. December to June of 2013).

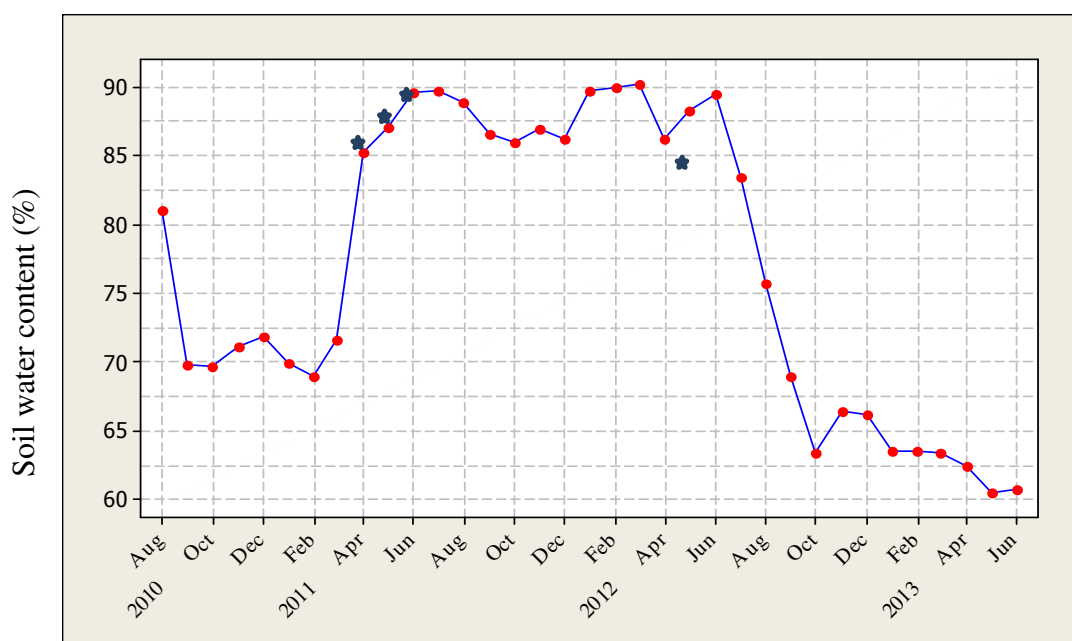


Figure 4.3. Changes in mean monthly soil water content (SWC) (%) from August of 2010 to June of 2013. ★ : Flooding period.

Mean monthly soil temperatures (T_s) are illustrated in Figure 4.4. It is clear that there was a decreasing trend from summer (June to September) to winter (December to March). The minus degree was occurred only in January of 2011. Mean T_s was the highest in July for both 2011 (19.9°C) and 2012 (18.4°C). Mean monthly air temperature (T_{air_mean}) was presented in Figure 4.5. There was similar trend in T_{air_mean} as with T_{s_mean} and the coefficient of determination (r^2) between them was 97.8%. The maximum T_{air_mean} for July of 2011 and 2012 was 18.7°C and 18.8°C, respectively, and the negative T_{air_mean} values were recorded in February of 2011 and 2012.

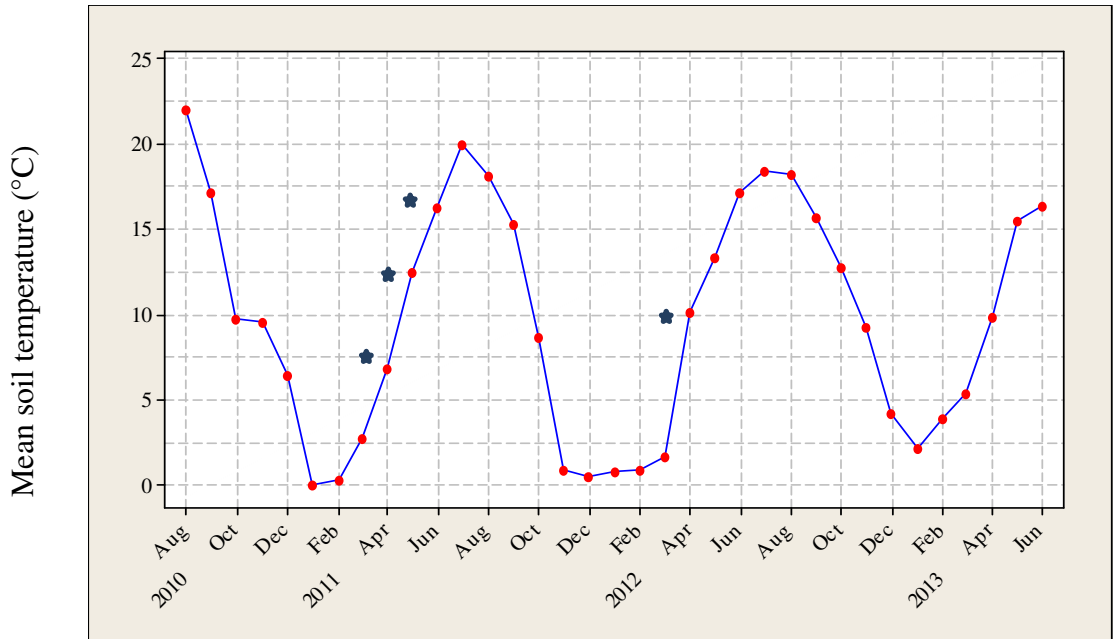


Figure 4.4. Changes in mean monthly soil temperature (T_{s_mean} , °C) from August of 2010 to May of 2013. ★ : Flooding period.

Water table level (WL) represents the vertical distance of groundwater from the surface peat layer. The WL measurements were performed from January of 2011 to June of 2013 but not a regularly basis (Figure 4.6). The flooding period was shown in Figure 4.6 as zero (April to June of 2011 and April of 2012). Due to the drought of

2012, WL from June to November of 2012 was higher than that of 2011.

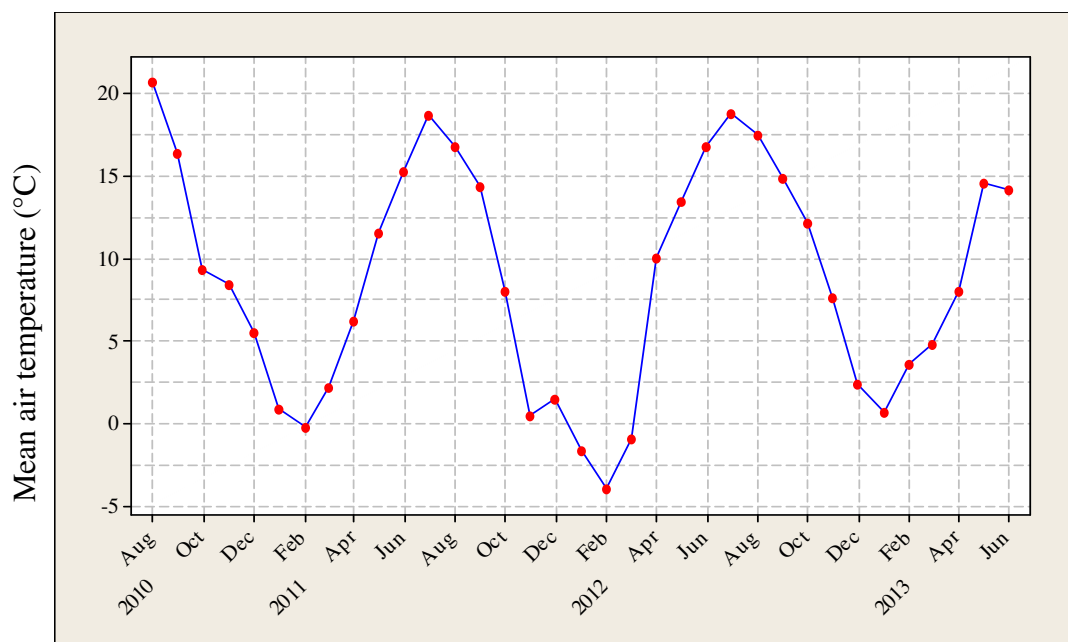


Figure 4.5. Changes in mean monthly air temperature ($T_{\text{air_mean}}$, °C) from August of 2010 to May of 2013.

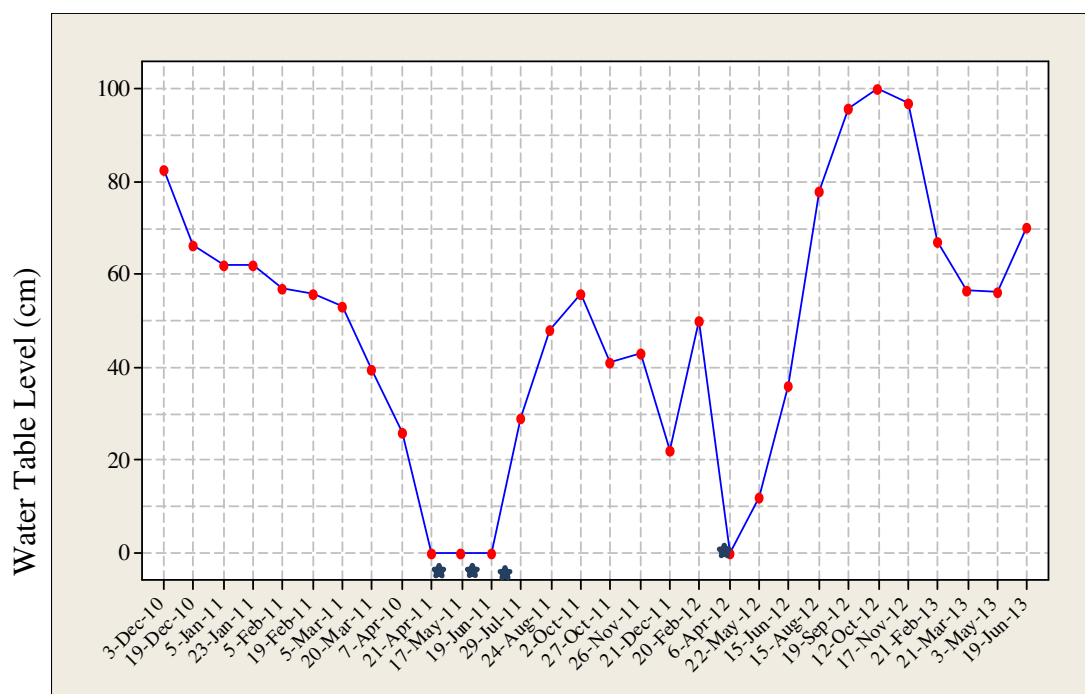


Figure 4.6. Changes in water table level (WL, cm) between 2010 and 2013.

★ : Flooding period.

4.2. Ancillary measurements

4.2.1. Lignin, carbon and nitrogen contents of peat soil and plant residues

Peat soil and plant residues in Yeniçağa peatland were investigated in terms of their lignin, carbon (C) and nitrogen (N) contents. The peat soil and plant residue samples were analyzed for 19 different points. Results for the sampling points are introduced in Table 4.1 with the mean and standard deviation (SD) values.

Table 4.1. Peat soil and plant residue properties at EC measurement site

Sample Number	Lignin (%)	Plant residue			Peat soil		
		C (%)	N (%)	(C:N)	C (%)	N (%)	(C:N)
1	6.57	40.9	2.8	14.4	22.0	1.8	12.3
2	7.75	39.4	2.5	15.8	24.5	1.8	13.8
3	12.37	38.8	1.5	26.6	25.4	1.9	13.1
4	9.59	38.8	1.7	23.1	22.4	1.8	12.3
5	3.97	39.3	2.8	13.9	35.0	1.7	21.1
6	7.14	37.4	2.6	14.5	12.1	1.1	11.0
7	13.98	40.6	1.3	32.3	30.2	2.0	15.2
8	5.38	39.0	2.6	15.0	27.6	1.8	15.7
9	6.78	37.4	2.2	17.3	25.4	2.0	13.0
10	6.06	39.4	4.4	9.0	3.7	0.4	8.6
11	4.90	36.1	4.7	7.7	27.3	1.8	15.2
12	3.96	43.2	3.9	11.0	14.0	1.0	13.5
13	10.42	41.6	1.2	35.4	35.1	2.2	16.1
14	6.67	41.4	2.5	16.3	15.5	1.3	11.6
15	4.08	39.8	4.2	9.4	21.6	1.7	12.6
16	4.84	39.3	2.6	15.2	31.5	2.2	14.2
17	8.58	41.8	1.1	37.3	28.8	2.2	13.1
18	6.76	40.8	1.7	23.6	19.2	1.5	12.8
19	7.45	41.9	2.2	19.1	20.8	1.5	13.9
Mean	7.2	39.8	2.6	18.8	23.3	1.7	13.6
SD	2.7	1.8	1.1	8.8	8.0	0.5	2.5

Mean C and N contents of peat soil were 23.3% and 1.7%, respectively. C content was lower and N content was higher than C and N results for Yeniçağa peat soil (35.2% C and 1.5% N contents) reported by Dengiz et al. (2009).

Monthly aboveground biomass was measured from June of 2012 to May of 2013. The mean and SD values of dry weight (DW) aboveground biomass at five sampling points were introduced in Table 4.2. DW aboveground biomass results were converted to g C per area (m^2) per month (month^{-1}) by multiplying the dry weight aboveground biomass weight with a conversion factor (0.5) (Evrendilek et al., 2006) (Figure 4.7). The mean C stock aboveground biomass was high in June and August of 2012 and decreased during the remaining period called hereafter as non-growing season.

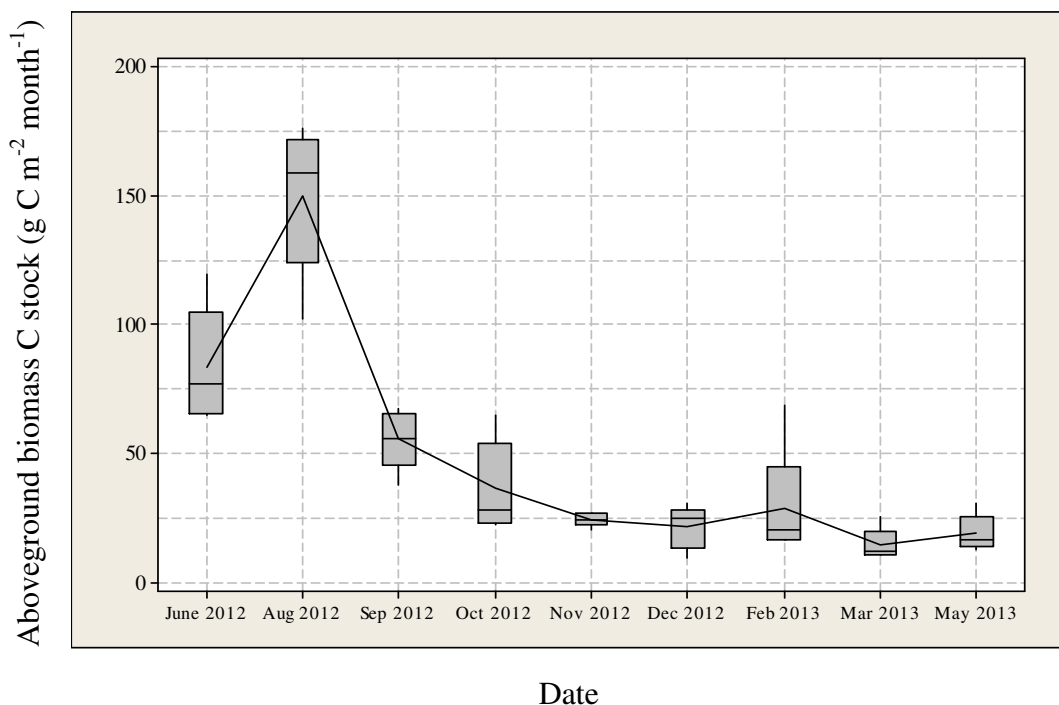


Figure 4.7. Changes in aboveground biomass C stock ($\text{g C m}^{-2} \text{ month}^{-1}$).

Litter quality was defined in different ways in the related literature such as N content, lignin content, and C:N ratio (Figure 4.8). Thus, the decomposition rate of litters can be predicted using C:N ratios, lignin content or lignin:N ratios. In general, litter decay rate is inversely related to C:N and lignin:N ratios. In other words, high quality litter (low C:N ratio) decomposes faster than low quality litter (high C:N ratio) (Manzoni et al., 2008; Zeng et al., 2009).

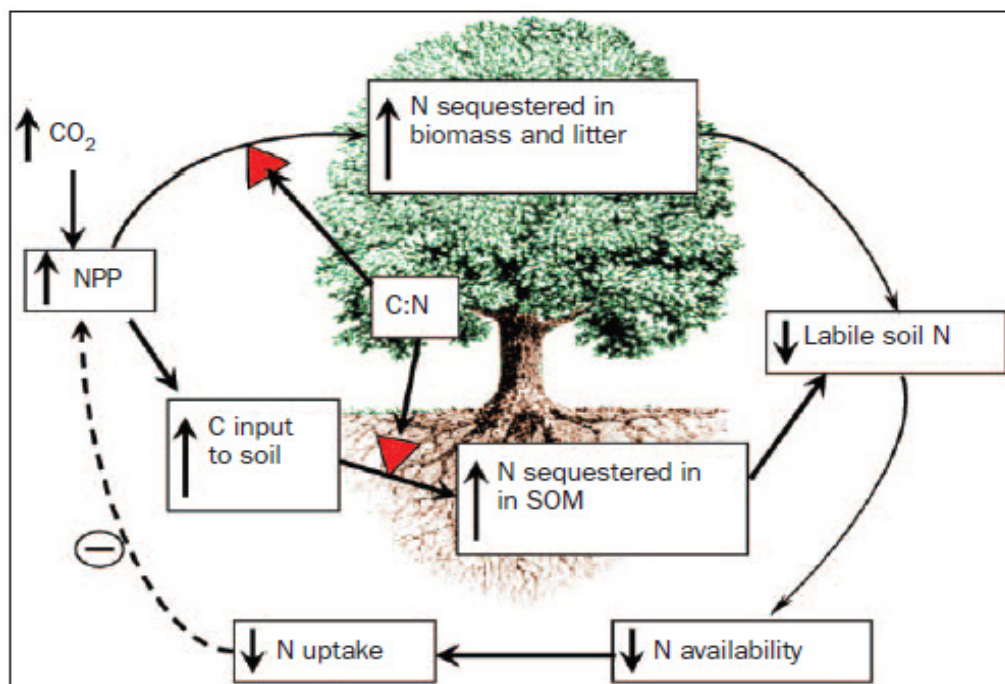


Figure 4.8. Two sets of feedback processes to elevated carbon dioxide (CO₂) leading to progressive nitrogen (N) limitation. In the upper pathway, the initial productivity response to elevated CO₂ results in N sequestration in plant biomass and litter pools, slowing N release to labile soil N. In the lower pathway, increased net primary production (NPP) in response to elevated CO₂ increases carbon (C) input to soil (increased exudation, root growth and death, and aboveground litterfall), leading to increased N sequestration in soil organic matter (SOM). This N sequestration reduces N availability to plants and subsequent plant N uptake. Flexible C:N ratios in ecosystem organic matter pools can modify both pathways. Flexible soil C:N pools could allow increased C storage without restricting short to medium-term N availability. A flexible plant C:N ratio could temporally decrease N demand (Luo et al., 2004).

Litter decay rate was determined with a litterbag technique (Moore et al., 2007). The meshed litterbags weighed initially were buried to a soil depth of 5 cm at five sampling points and were taken out of their places periodically at six times to weigh remaining mass. Remaining DW litter mass of the mesh litterbags and their C and N contents are introduced in Table 4.2. The results of the DW litter mass remained after days 60, 120, 160, 200, 240 and 280 were given in the form of g C m^{-2} unit by multiplying with 0.5 to obtain C content (Evrendilek et al., 2006) and dividing to 0.475 m^2 ($19 \times 25 \text{ cm}$ litterbag area) to obtain C content per m^2 (Figure 4.9). The mass loss decreased exponentially with time. An exponential function was fitted into the data as follows:

$$\text{Remaining mass of dry weight litter (g C m}^{-2}\text{)} = 26.95 * e^{-0.006\text{time}} \quad (r^2 = 0.94; n = 7; p < 0.05).$$

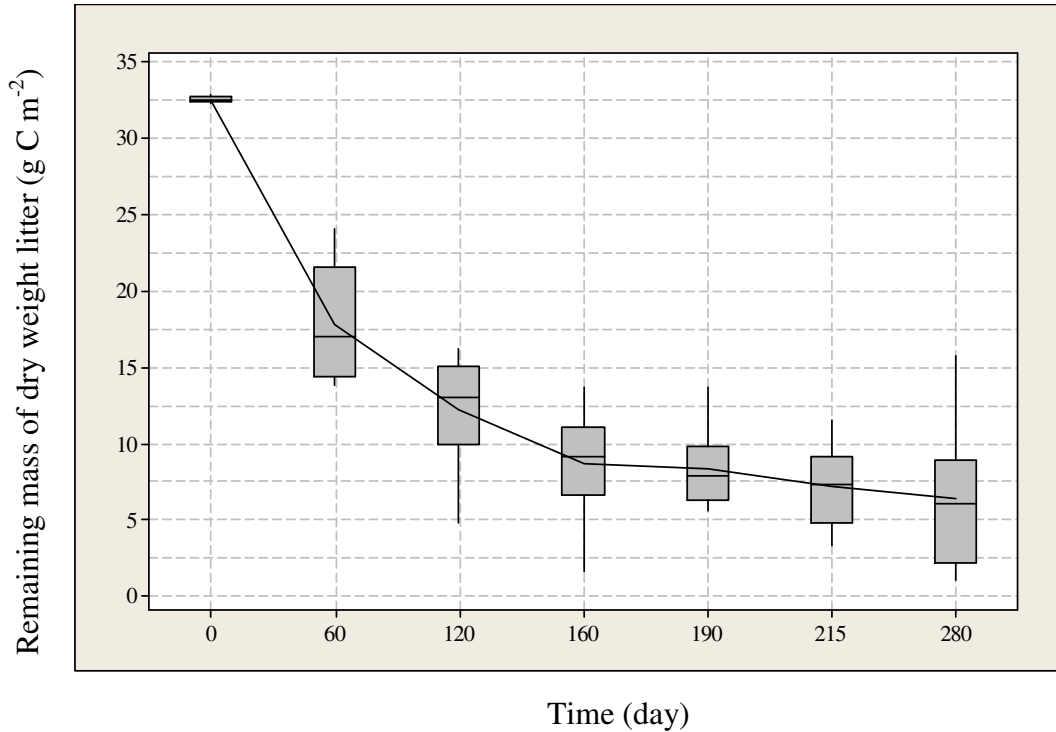


Figure 4.9. Change in remaining litter mass (g C m^{-2}) over time (day). Initial day of year for the litter meshed litterbag experiment = 24/06/2012; and the initial litter dried mass = 3.09 g C m^{-2} .

Results for litter quality (C:N = 18.8) and decomposition constant (-0.006) indicated that the decomposition rate of SOM in Yeniçağa was exerting a positive effect on CO₂ efflux (e.g. Moore 2007; Anderson, 2008; Manzoni et al., 2010; Broder et al., 2012).

Table 4.2. Peat soil decomposition rate and aboveground biomass stock

	Litter decay rate in peat soil ($\text{g C m}^{-2} \text{ day}^{-1}$)									Aboveground biomass ($\text{g C m}^{-2} \text{ month}^{-1}$)								
Sample Number	N (%)	C (%)	$t = 0$ June 2012	$t = 60$ Aug 2012	$t = 120$ Oct 2012	$t = 160$ Nov 2012	$t = 190$ Dec 2012	$t = 215$ Feb 2013	$t = 280$ May 2013	June 2012	Aug 2012	Sep 2012	Oct 2012	Nov 2012	Dec 2012	Feb 2013	March 2013	May 2013
1	1.7	38.8	32.7	24.1	13.6	10.3	13.7	7.5	6.6	65.1	176.4	67.7	42.7	26.9	30.9	21.1	11.1	15.1
2	2.1	40.9	32.5	20.7	16.2	9.7	7.5	11.6	15.7	65.5	102.4	37.7	22.4	20.7	16.6	16.5	12.1	20.7
3	2.4	36.1	32.3	13.9	4.8	1.7	5.7	3.4	1.1	90.7	159.1	63.5	64.9	24.6	25.1	16.3	11.1	12.6
4	2.2	39.4	32.4	16.1	14.7	13.7	8.5	8.4	5.8	119.5	146	55.9	28.3	23.9	25.3	20.6	14.1	16.6
5	2.5	37.4	32.9	14.6	11.7	8.3	6.5	5.4	2.6	77.4	167.1	53.2	23.6	27.2	9.7	68.9	25.7	30.7
Mean (g C m^{-2})	2.2	38.5	32.6	17.9	12.2	8.7	8.4	7.2	6.4	83.6	150	55.6	36.3	24.7	21.5	28.7	14.8	19.1
SD	0.3	1.8	0.3	4.4	4.4	4.4	3.2	3.1	5.7	22.6	28.9	11.6	17.9	2.7	8.3	22.6	6.2	7.1

SD: standard deviation

4.3. Carbon dioxide fluxes

4.3.1. Data analysis

The processing and filtering of EC data is an important step to achieve reliable flux data. The strategy of the data processing and filtering was mentioned in Section 2.4.3. In this study, the processing and filtering strategy was decided based on the information described in Section 2.4.3. The strategy is illustrated in Figure 4.10.

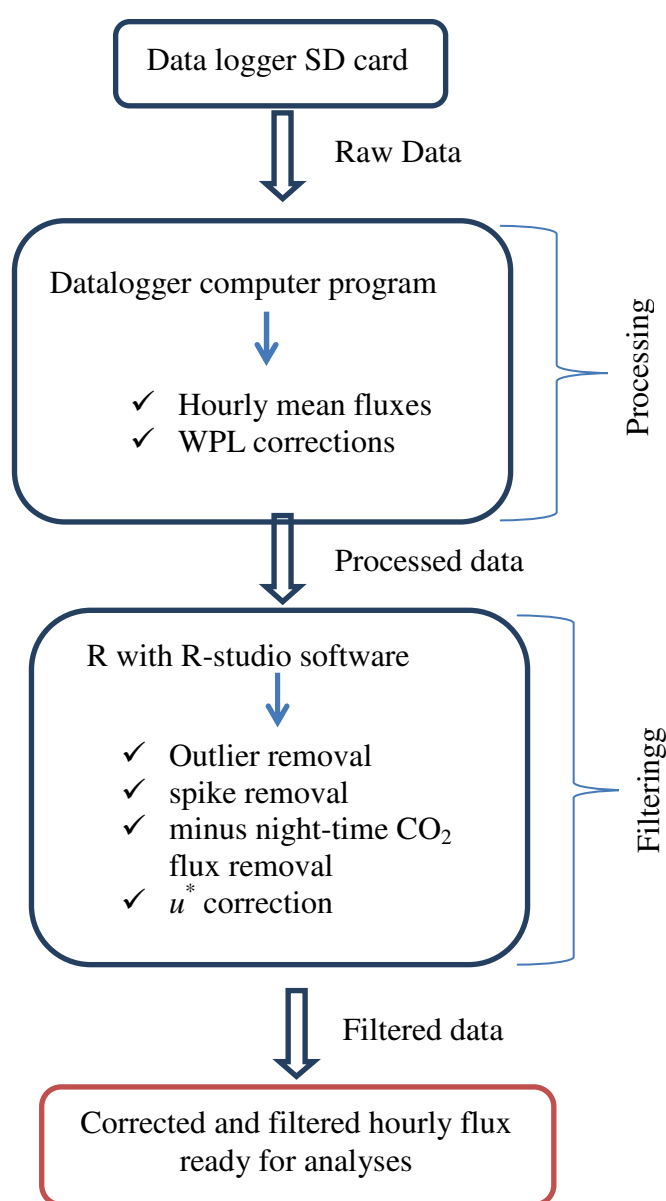


Figure 4.10. A schematic illustration of raw data processing and filtration steps.

Data logger (CR300 Campbell Scientific Inc.) collects raw data and hourly mean fluxes data. The first preprocessing is done by the data logger (Campel CR300) with the application of WPL-correction (Webb, Pearman and Leuning 1980, WPL). This correction is required because of fluctuations in temperature and humidity in the atmosphere which in turn affect the concentration fluctuation of the trace gases (Aubinet et al., 2012). Basically, the infrared gas analyzer (Licor-7500 open) is used for CO₂ molar density. The molar density of CO₂ (or the gases to be detected) means the number of molecules in sampled air volume. The air sample contains many molecules and air density can be affected by variations in temperature and humidity (Aubinet et al., 2012). For example, CO₂ density of a dry-air sample is higher than that of a moist-air sample due to the decreased volume of the dry air. Thus, a correction should be done during the measurement of CO₂ fluxes in moist air. The application of the correction should be performed carefully; otherwise, many of the available raw flux data can be removed unnecessarily. The second step is to apply a filtering procedure to processed data. Threshold values for filtering are site-specific (Moffat et al., 2007; Baldocchi, 2008; Aubinet et al., 2012). Discarding of outliers was followed by rejection of spikes, separation into daytime and nighttime fluxes, and filtering by friction velocity (u^*), (Thomas et al., 2011; Liu et al., 2012).

The processed data (hourly mean fluxes data with WPL-corrected) are recorded in TOA5 format by the data logger. The processed data (~30000 rows) were imported into Microsoft Excel (2010) files. The detection of outliers is time-consuming and error-prone. R with R-studio is one of the software programs available to be used for that purpose. The elimination procedure should be applied carefully to minimize the risk of removing meaningful data. CO₂ density fluctuations and variations in latent

and sensible heat fluxes in response to CO₂ fluxes should be taken in to account. The procedure about how to proceed with this purpose is explained below.

Standard deviation of CO₂ density (SD_{CO₂}) versus CO₂ flux (F_{CO₂}) was plotted to investigate this relationship. As can be seen in Figure 4.11, there is a data aggregation when SD_{CO₂} values vary between 0 and 200 which indicates that CO₂ density fluctuation in this range (0 - 200) does not affect F_{CO₂} measurements. However when SD_{CO₂} is higher than 200, F_{CO₂} data get higher without a regular pattern observed (Figure 4.11) Thus, F_{CO₂} data were removed when SD_{CO₂} value is higher than 200.

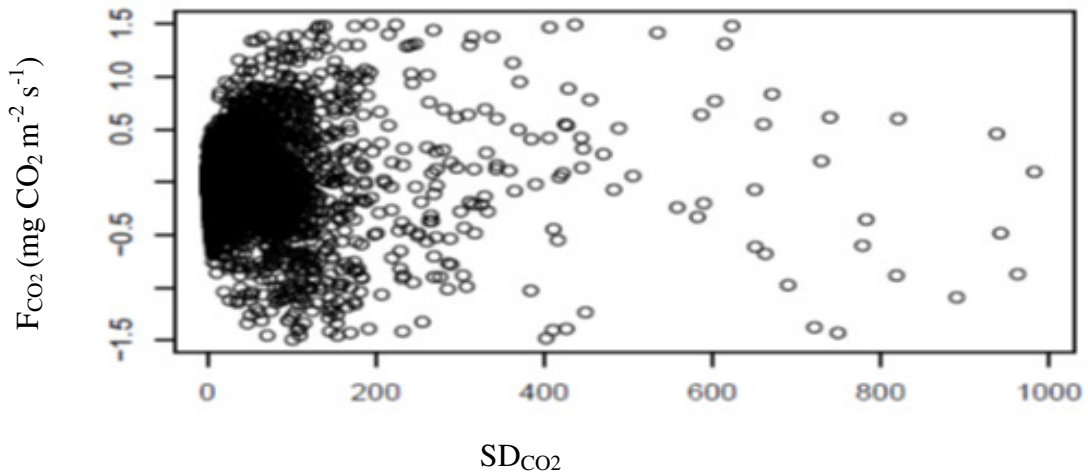


Figure 4.11. A plot of CO₂ flux (F_{CO₂}) as a function of standard deviation in CO₂ density (SD_{CO₂}).

Latent heat (LE) and sensible heat (H) fluxes are the other important variables that can affect the precision of CO₂ flux data. LE and H fluxes ranged from -200 to 600, and the range of -100 to 250 was selected according to Figures 4.12 and 4.13, respectively.

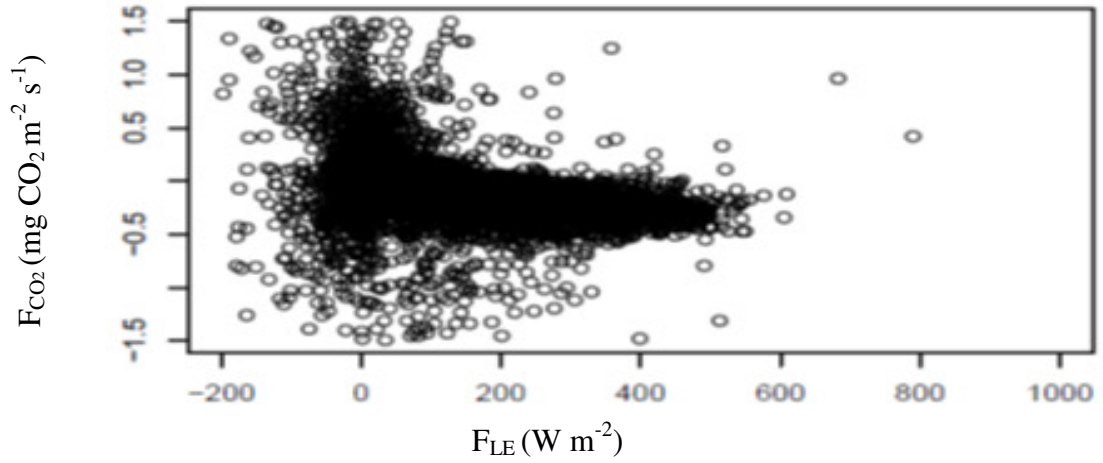


Figure 4.12. A plot of CO₂ flux (F_{CO_2}) as a function of latent heat flux (F_{LE}).

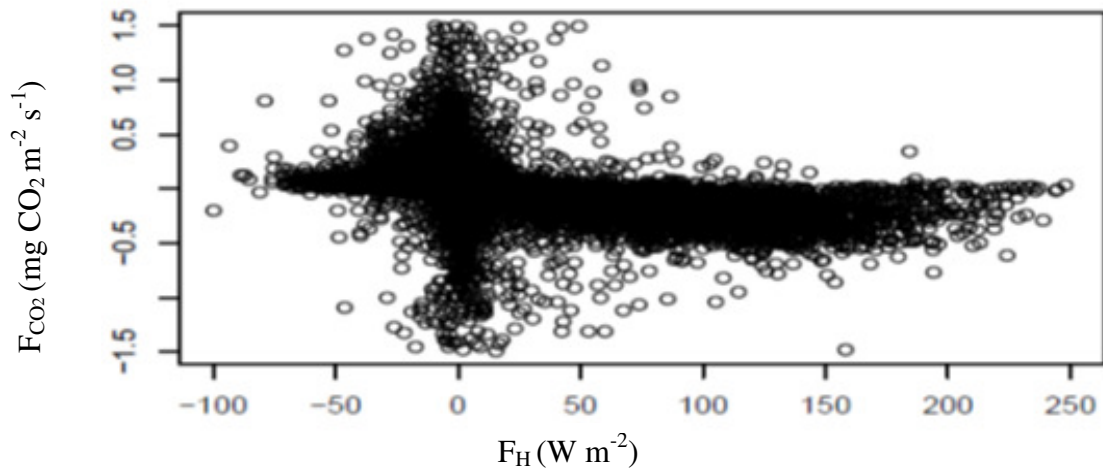


Figure 4.13. A plot of CO₂ flux (F_{CO_2}) as a function of sensible heat flux (F_H).

After the removal of outliers, variations in CO₂, LE and H fluxes over time were plotted. According to Figure 4.14, there are still some spikes which should be further removed. The spike data were defined as three times x standard deviation plus mean. The time window was selected as three weeks (with weeks before and after) to calculate mean fluxes and standard deviations for determining a threshold value. If the data are higher than calculated threshold value, then the flux is removed. This algorithm was run for each year. This screening procedure was applied to CO₂, LE

and H fluxes. Another filtering is applied to CO₂ flux data when nighttime fluxes are negative which indicates CO₂ uptake by the ecosystem. After the application of this filtering procedure, time-series graphs were plotted for each of CO₂, LE and H fluxes as shown in Figures 4.15 to 4.17.

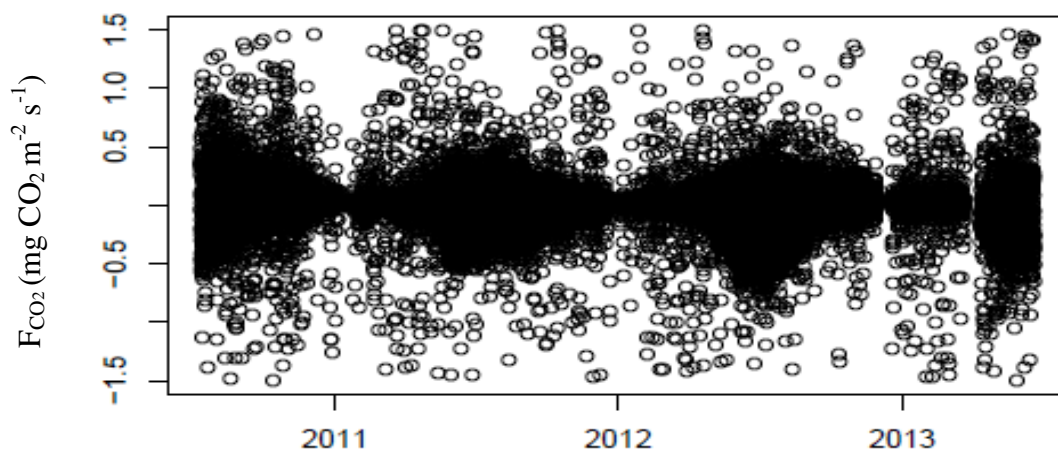


Figure 4.14. CO₂ flux (F_{CO_2}) variations between 2010 and 2013 before filtering of data.

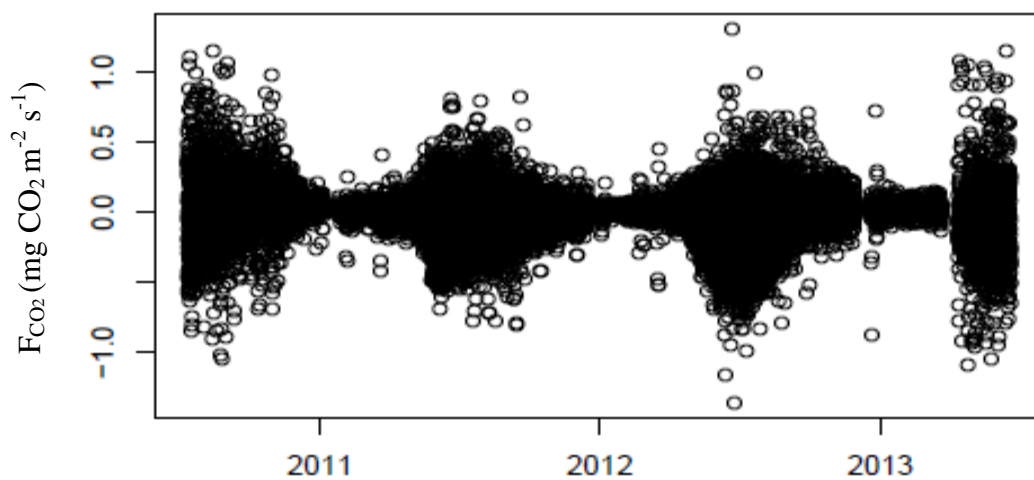


Figure 4.15. CO₂ flux (F_{CO_2}) variations between 2010 and 2013 after filtering of data.

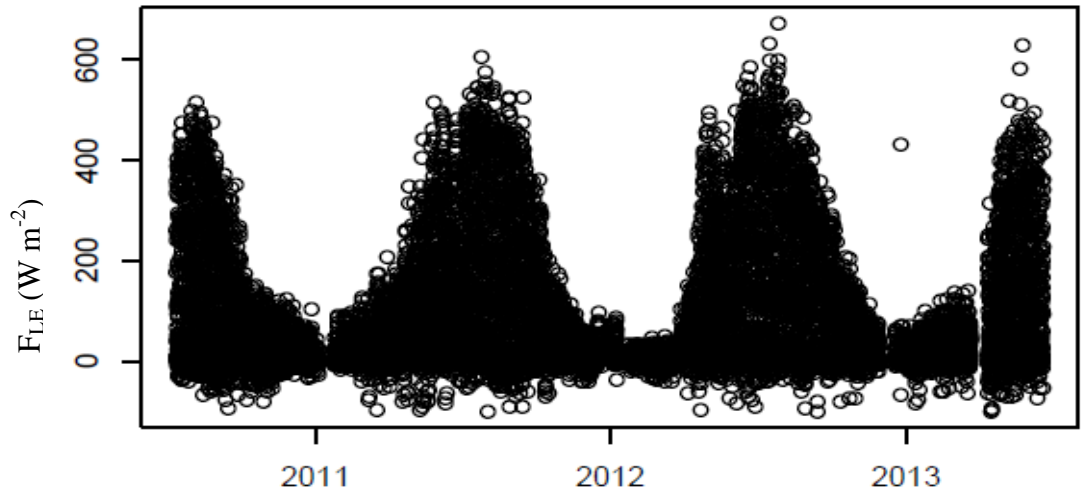


Figure 4.16. Latent heat flux (F_{LE}) variations between 2010 and 2013 after filtering of data.

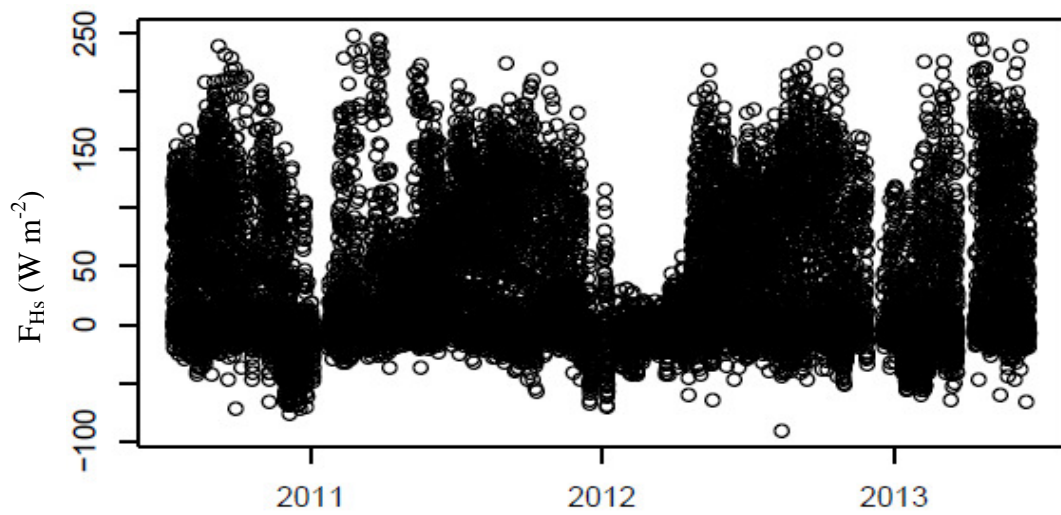


Figure 4.17. Sensible heat flux (F_H) variations between 2010 and 2013 after filtering of data.

The next step was to separate the dataset into daytime and nighttime fluxes. The flux data were partitioned into daytime and nighttime according to a global radiation threshold of 20 W m^{-2} (Reichstein et al., 2005; Wilkinson et al., 2012; Polsenaere et al., 2012). The EC technique often underestimates nighttime NEE under stably stratified atmospheric conditions (eg. Black et al., 2000; Valentini et al., 2000). Therefore, this problem is still under discussion (Finnigan, 1999; Baldocchi et al.,

2001). The stable atmospheric conditions typical at night tend to suppress turbulence, thus causing a decrease in atmospheric mixing. In these conditions, the trace gas can accumulate close to the ground not reaching the sensor height and causing night time gas flux underestimations (Baldocchi, 2003). The practical solution used in the literature is to filter the nighttime measurements due to the atmospheric turbulence conditions using friction velocity (u^*) as an indicator.

The filtering procedure applied for u^* selection was described by Goulden et al. (1996). It was suggested that a threshold for u^* be defined and the flux data be discarded below that threshold value. This threshold is determined plotting nighttime CO_2 fluxes as a function of u^* data to better see this relation. When the flux is not dependent on the u^* , u^* critical value is defined as a threshold. For this study, no clear correlation was found between the friction velocity (u^*) and the night CO_2 fluxes in Figure 4.18, and therefore, u^* filtering was not performed.

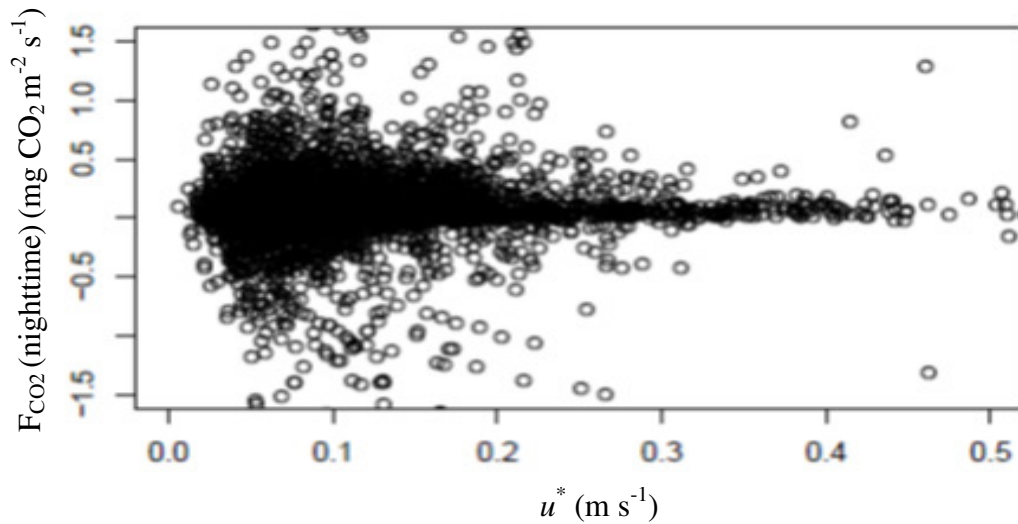


Figure 4.18. Selection of u^* threshold.

4.3.2. Gap-filling used in the study

The unfavorable conditions (e.g. high rainfall) especially in the non-growing season, and the filtering procedure gave rise to a higher percentage of gap-filling than in the growing season. The gap-filling methods are introduced in Section 2.4.4. In this study, the missing values were substituted with estimates using a standard gap-filling procedure employed by CarboEurope (Reichstein et al., 2005). This Marginal Distribution Sampling (MDS) method is effectively an enhancement of look-up tables (LUT) whereby flux measurements are correlated to meteorological conditions and produced consistently accurate results when compared to the other more complex methods (Section 2.4.4) (Moffat et al., 2007). Therefore, this method has been used extensively by many studies (e.g. Thomas et al., 2011; Marushchak et al., 2012; Wilkinson et al., 2012; Zona et al., 2013; Morrison et al., 2013). The basic requirements for MDS method include CO_2 , H and LE flux data and the meteorological variables of solar radiation (R_g), air temperature (T_{air}) and vapour pressure deficient (VPD). The MDS method requires the following three conditions:

- 1: all meteorological data are available. In that situation, the missing data are replaced by the average value under similar meteorological conditions within a time-window of seven days. Where the data gaps are large the averaging window can be increased from seven days, but these data are then assigned to a lower accuracy category.
- 2: if T_{air} and VPD are missing, while only R_g is available, then similar R_g conditions are taken into account where the SD values of R_g are less than 50 W m^{-2} .
- 3: if all meteorological data are missing, then the missing value is replaced by the average value at the same time of the day (1 hour), i.e. by the mean diurnal variation (MDV, Section 2.4.4.).

The number of gap-filled data for daytime and nighttime period on a monthly basis during the measurement period are introduced in Table 4.3.

Table 4.3. Percentage of the gap-filled data for daytime and nighttime on a monthly basis in 2010, 2011, 2012 and 2013

Months	Daytime Data	Nighttime Data	Gap-filled daytime data % of daytime data				Gap-filled nighttime data % of nighttime data			
			2010	2011	2012	2013	2010	2011	2012	2013
January	403	341	NA	80.4	70.7	43.9	NA	83.8	75.9	53.4
February	364	308		45.6	44.8	28.6		59.4	49.2	35.4
March	403	341		37.2	45.2	51.9		56.6	65.9	58.1
April	390	330		30	30	51		59.4	44.5	63.6
May	403	341		30.8	21.6	14.4		33.7	53.4	47.5
June	390	330		17.2	16.4	25.2		32.1	55.3	46.8
July	403	341		11.2	18.1	NA		39.3	57.8	NA
August	403	341	6.5	15.1	24.3		30.4	55.7	59.2	
September	390	330	8.5	25.1	28.9		90	217	227	
October	403	341	31.8	28.3	27.1		38.1	47.2	60.7	
November	390	330	16.4	28.9	34.3		34.5	59	63.8	
December	403	341	51.4	65.3	63.2	56.9	71.8	70.2		

*NA: Not available, the measurements could not be performed during that period.

The percentage of the gap-filled data in nighttime was generally higher than in daytime. As mentioned earlier, this was due to low turbulence during the nighttime period that prevents flux measurement by EC sensors (Aubinet et al., 2012). In addition, while the percentage of gap-filled data was higher in winter and lower in summer because of unfavorable winter conditions (e.g. rainfall). The mean percentage of gap-filled data was 28.7%, 34.6%, 35.4%, 35.8% for daytime and 37.4%, 55.3%, 60.4%, 50.8% for nighttime in 2010, 2011, 2012, and 2013, respectively. The mean percentages of daytime and nighttime gap-filled data are similar with those of the related literature (between 20% and 69%) (Moffat et al., 2007).

4.3.3. Daily NEE flux

Hourly and mean daily NEE fluxes for the entire measurement period were estimated using gap-filled data. NEE flux was minus for the daytime period between 07:00 and 20:00 and positive for the nighttime periods between 21:00 and 06:00 (Figure 4.19). In other words, the Yeniçağa peatland acted as a sink during daytime period and acted as a source during the nighttime period.

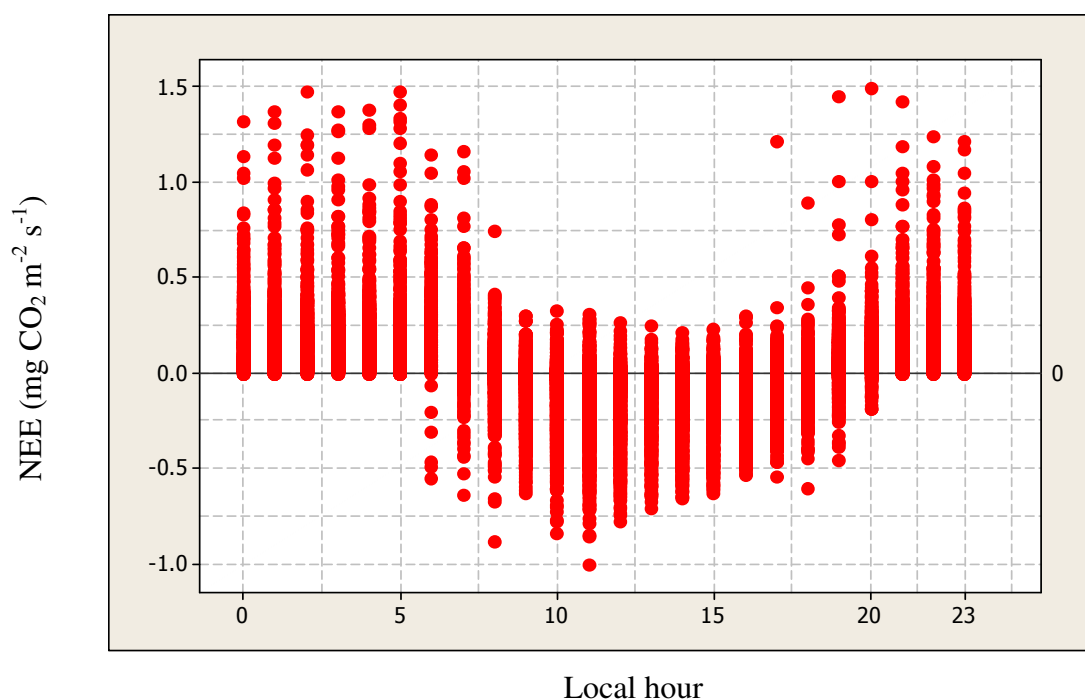


Figure 4.19. Hourly net ecosystem exchange (NEE) measurements via the EC system for Yeniçağa peatland for the entire monitoring period. Negative values imply CO₂ uptake by the ecosystem.

The mean daily NEE plot from August of 2010 to June of 2013 was shown in Figure 4.20. The maximum daily uptake ($\sim -0.16 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) occurred between days of year (DOY) 152 in 2011 and DOY 213 in 2012. CO₂ emissions occurred between DOY 213 and 274 ($\sim 2.7 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$).

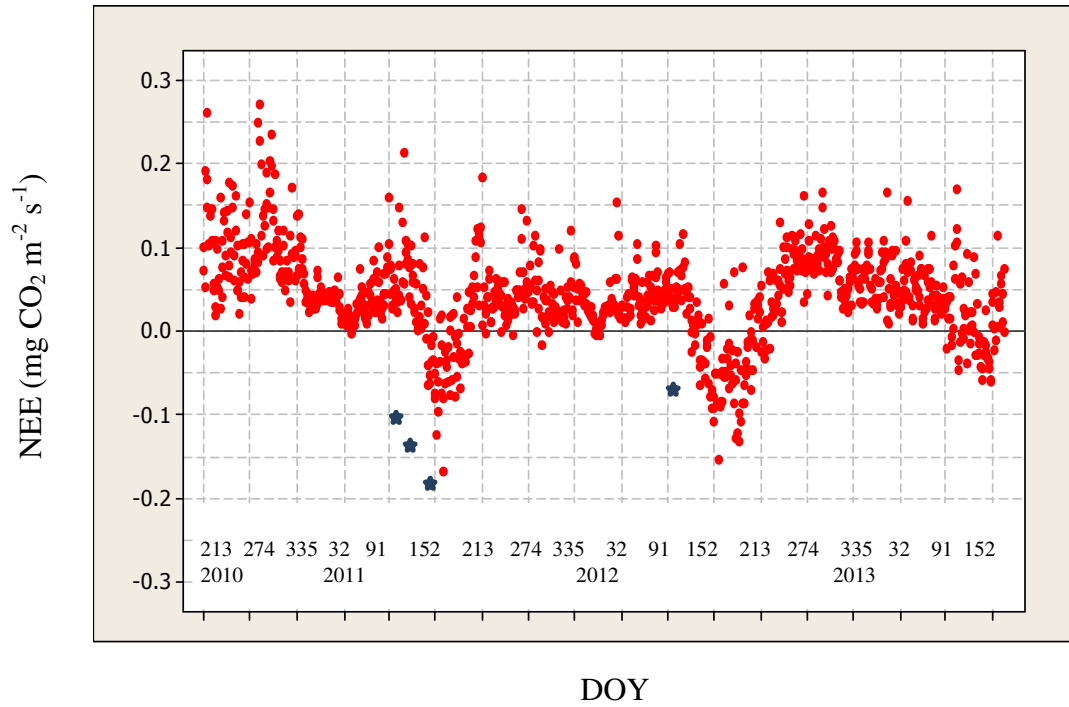


Figure 4.20. Mean daily NEE measurements via the EC system for Yeniçağa peatland between 01 August 2010 and 19 June 2013. ★ : Flooding period (April to June of 2011 and April of 2012).

The net mean daily NEE values were 0.097, 0.033, 0.031 and 0.036 mg CO₂ m⁻² s⁻¹ for 2010, 2011, 2012 and 2013, respectively (Table 4.4). The mean daily NEE was higher in 2010 than in the others because 2010 data consisted of only fall and winter measurements (August to December). The mean daily CO₂ concentrations in the atmosphere recorded by EC system during the measurement period (August of 2010 to June of 2013) are shown in Figure 4.21. The mean daily CO₂ concentration was observed as 314.3 ± 28.8 ppm (Table 4.5).

Table 4.4. Descriptive statistics for daily NEE measurements ($\text{mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) in 2010, 2011, 2012, and 2013

NEE ($\text{CO}_2 \text{ mg m}^{-2} \text{ s}^{-1}$)	<i>N</i>	Mean	SD	CV	Min	Median	Max
2010	153	0.097	0.052	53.2	0.019	0.089	0.272
2011	365	0.033	0.044	134.3	-0.167	0.034	0.215
2012	349	0.033	0.055	166.8	-0.154	0.038	0.167
2013	170	0.036	0.043	118.0	-0.059	0.036	0.170

N: number of measurement days; SD: Standard deviation; CV: Coefficient of variance; Min: Minimum; Max: Maximum

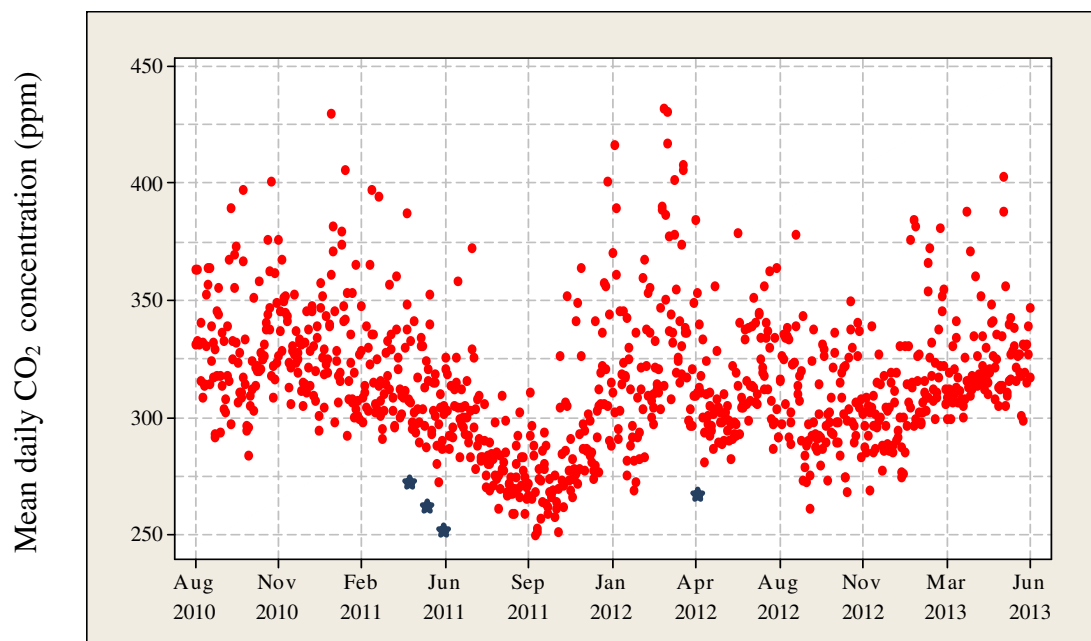


Figure 4.21. Mean daily atmospheric CO_2 concentration (ppm by volume) measured via the EC system for Yeniçağa peatland between August of 2010 and June of 2013.

★ : Flooding period (April to June of 2011 and April of 2012).

Table 4.5. Descriptive statistics for atmospheric CO_2 concentrations between August of 2010 and June of 2013

Variable	<i>N</i>	Mean	SD	CV	Min	Median	Max
Atmospheric CO_2 (ppm)	1021	314.3	28.8	9.2	151.9	312.1	432.1

N: number of measurement days; SD: Standard deviation; CV: Coefficient of variance; Min: Minimum; Max: Maximum

4.3.4. Monthly NEE flux

Mean monthly NEE data was calculated in order to investigate monthly variation (Figure 4.22). The Yeniçağa peatland was a source for the measurement period of 2010, a sink for only June of 2011, a sink from May to July of 2012 and a sink for only May of 2013.

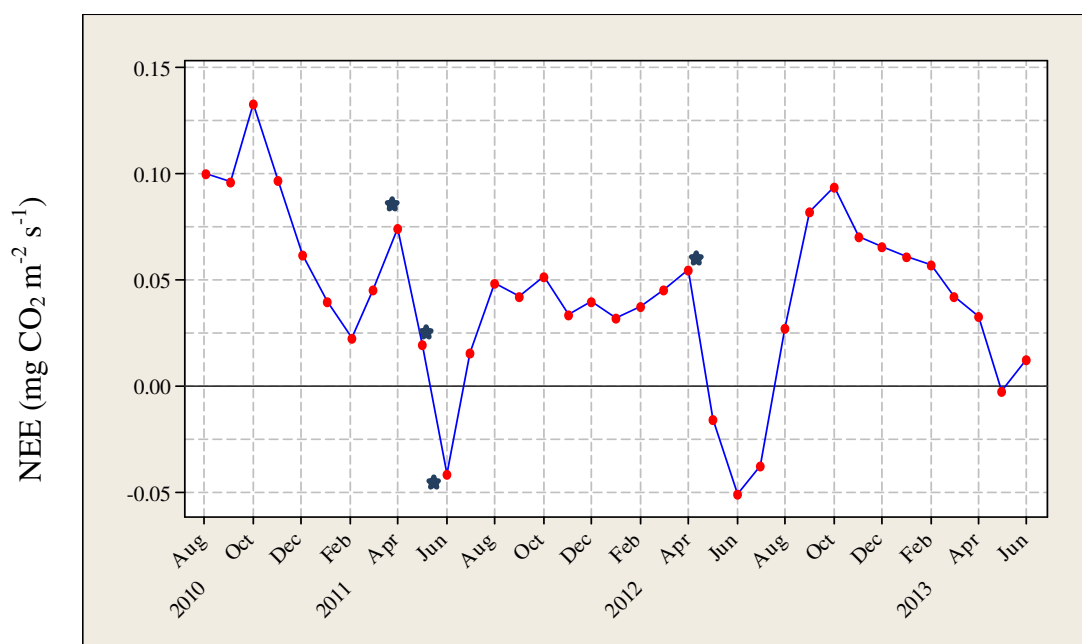


Figure 4.22. Mean monthly NEE measurements via the EC system for Yeniçağa peatland from 01 August 2010 to 19 June 2013. ★ : Flooding period.

The highest CO₂ emissions to the atmosphere occurred for October (2010 and 2013) and April (2011 and 2012). The highest CO₂ uptake occurred in June of 2011 and 2012. The source strength decreases from April to May, then begins to act as a sink in June. Subsequently the source strength increases from July to October in 2011 and 2012.

4.3.5. Annual NEE flux

The missing data were filled with a MDS gap-filling and provided complete dataset for the measurement period (August of 2010 to June of 2013). The hourly NEE data varied between -1.0 and 1.5 CO₂ mg m⁻² s⁻¹ for the measurement period (Figure 4.23). The annual cumulative CO₂ balance of the Yeniçağa peatland was estimated for the measured period at 352.7, 283.8, 252.4 and 142.6 g C m⁻² in 2010, 2011, 2012 and 2013, respectively (Table 4.6). The annual cumulative NEE was highest in 2010 and lowest in 2013. The reason for that is the incomplete NEE measurements of the years. However, the annual cumulative NEE values were similar in 2011 and 2012 which showed the trend for the complete year measurements. The seasonal variation showed that the source strength was lowest in summer for both 2011 and 2012, whereas, the highest source strengths varied seasonally for the both years. The highest source strength was obtained in winter of 2011 and in fall of 2012 (Table 4.6). Once cumulative NEE was seasonally compared for 2011 and 2012; the source strength decreased in the following order: winter > spring > fall > summer in 2011, and fall > winter > spring > summer in 2012.

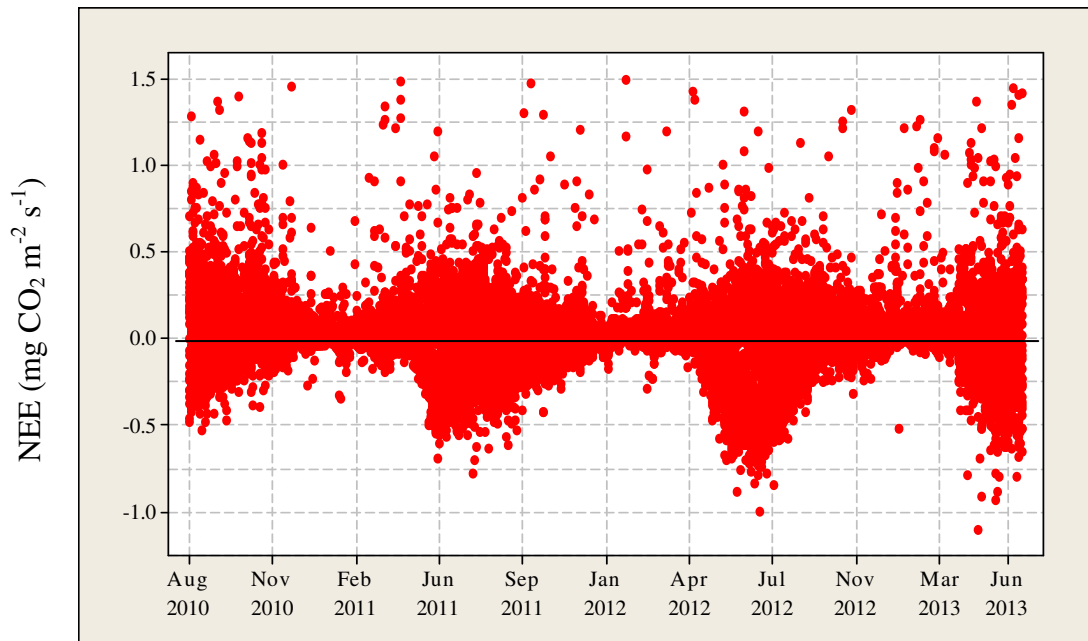


Figure 4.23. Annual NEE trends at Yeniçağa peatland. Points refer to one hour averages of measured data.

Table 4.6. Summary of seasonal and annual NEE on an annual basis^a

Period	2010	2011	2012	2013
Spring	NA	252.4	105.9	80.79
Summer	519.7 (only August-September)	173.9	56.3	18.5 (only June)
Fall	607.7	226.1	434.7	NA
Winter	165.7 (only December)	387.9	328.5	423.4 (only January-February-March)
Annual	1293.1	1040.3	925.4	522.7
g C m ⁻² period ⁻¹	352.7	283.8	252.4	142.6

^a Annual period is from 1 August to 31 December for 2010, from 1 January to 31 December for 2011 and from 1 January to 28 November for 2012. Seasonal periods are Spring (April - May), Summer (June-July-August-September), Fall (October-November), and Winter (December-January- February-March). All values are NEE in g CO₂ m⁻² for the periods defined unless otherwise stated. NA: not measured.

4.3.6. Components of carbon budget

The carbon balance is composed of the following three main fluxes; NEE, gross primary production (GPP) and total ecosystem respiration (R_{eco}). The relationship among the three fluxes which is discussed in Section 2.2 is given in the following Eq. (2.4) (Aubinet et al., 2012):

$$NEE = - GPP + R_{eco} \quad (2.4)$$

The total ecosystem respiration (autotrophic and heterotrophic respiration) means releasing of CO_2 from ecosystem to atmosphere, positive as a conventional meteorological sign notation. CO_2 fluxes fixed by photosynthesis in an ecosystem are negative that denotes the flux movement from atmosphere to ecosystem. According to the mass balance conservation, the summation of $- GPP$ and R_{eco} gives NEE.

The eddy covariance (EC) technique measures NEE and does not give any information about how much of CO_2 flux is used in photosynthesis or in respiration process. Since GPP and R_{eco} fluxes are important for understanding the complete mechanism and necessary inputs for modeling ecosystem C dynamics, measured NEE fluxes should be partitioned into to R_{eco} and GPP flux components.

Generally, two different approaches are defined for the partitioning of NEE into GPP and R_{eco} (Reichstein et al., 2005). One of the flux partitioning algorithms is based on nighttime flux data and the other one is based on both daytime and nighttime flux data for the estimation of R_{eco} (Reichstein et al., 2005).

The nighttime data-based models rely on modeling R_{eco} according to soil or air temperature (Aubinet et al., 2012). Nighttime NEE CO₂ flux data (measured by EC) are assumed to equal to R_{eco} (because no photosynthesis occurs at night then, $NEE_{night} = R_{eco}$) (Aubinet et al., 2012). Thus, the nighttime NEE flux is firstly filtered from bad quality data (according to data filtering mentioned in Section 2.4.3). Then the filtered nighttime NEE flux data are taken as $R_{eco\ night}$ flux data and fitted into an ecosystem respiration model. There are two common ecosystem respiration models that are applied in flux partitioning algorithm as follows:

Q_{10} Model (Ryan 1991)

$$R_{eco} = R_{10} Q_{10}^{\left(\frac{T-T_0}{10}\right)} \quad (4.1)$$

where T: air or soil temperature (°C); R_{10} : ecosystem respiration at 10°C; Q_{10} : temperature sensitivity parameter.

After running regression analysis between R_{eco} (nighttime) and temperature $(T - T_0)/10$, R_0 and Q_{10} constants are determined.

Lloyd and Taylor (1994) Model

$$NEE_{night} = R_{eco_{night}} = R_{ref} e^{E_0 \left(\frac{1}{T_{ref} - T_0} - \frac{1}{T - T_0} \right)} \quad (4.2)$$

where E_0 : the activation energy parameter; R_{ref} : the ecosystem respiration at T_{ref} ; T_{ref} : the reference temperature is set to 10°C; T: air or soil temperature (°C); T_0 : kept constant at -46.02°C.

The regression analysis performs for $R_{eco\ night}$ and $e^{E_0\left(\frac{1}{T_{ref}-T_0}-\frac{1}{T-T_0}\right)}$ then, E_0 and R_{ref} can be observed.

These two R_{eco} models are very simple models that can be applied to flux data directly. This gives the advantage that ecosystem respiration can be determined directly with a simple application. However, nighttime-based models have some disadvantages that cause biases in the models. The nighttime flux data accuracy is very important for both model applications, and thus, filtering of bad nighttime data is necessary (Aubinet et al., 2012). The models only as a function of temperature mean no other factors than temperature affect R_{eco} significantly (Aubinet et al., 2012). Other factors aside from temperature such as soil moisture could influence R_{eco} (Reichstein et al., 2002). The addition of another explanatory variable into R_{eco} model allows for seasonally varying temperature sensitivity but the models based on site specific factors are not generic (Reichstein et al., 2005).

One approach for R_{eco} was to improve the Loyd and Tylor (1994) model based on temporally-varying (dynamic) estimation of R_{ref} (Reichstein et al., 2005). The E_0 was determined and held constant over the entire year. R_{ref} was determined according to a defined time interval (e.g. 30-day) (Falge et al., 2002; Law et al., 2002). This model led to a dynamic R_{eco} function; however, the extrapolation of the model to daytime resulted in a systematic error (bias) because long-term E_0 values may not reflect short-term responses (Reichstein et al., 2005).

A recent approach for R_{eco} adopted by Reichstein et al. (2005) was to further improve the Loyd and Tylor (1994) model in that E_0 was derived from short-term data. Thus,

the systematic error from the extrapolation to the daytime was minimized. This model has the following limitation: statistical significance of E_0 values decreases due to noises associated with long-term EC data.

An alternative to nighttime-based model for R_{eco} estimation is daytime- based models (Lasslop et al., 2010). In this approach the $NEE_{daytime}$ was modeled using the common rectangular hyperbolic light-response curve (Falge et al., 2001, Suyker and Verna 2001). Once the hyperbolic light-response curve (Eq. 3) is fitted for $NEE_{daytime}$ the y-intercept is equal to R_{eco} daytime flux. Different daytime-based R_{eco} models are described and discussed in the literature (Aubinet et al., 2012; Lasslop et al., 2010).

$$NEE_{daytime} = \frac{a' Q_{PPFD} GPP_{sat}}{GPP_{sat} + a' Q_{PPFD}} - R_{ecodaytime} \quad (4.3)$$

where a' : the initial slope; Q_{PPFD} : photosynthetically active photon flux density; GPP_{sat} : gross primary productivity flux at light saturation; $R_{eco \text{ daytime}}$: total ecosystem respiration at daytime ($R_{eco \text{ daytime}}$).

There is no standard strategy established for flux partitioning algorithm. The models mentioned above can be selected according to site properties, data quality and the model responses. In this study, nighttime-based model derived by Reichstein et al. (2005) was used for flux partitioning. Although this model application is not simple, this model is more reliable than the simple model owing to its advantage of minimizing systematic error.

Reichstein et al. (2005) improved the Lloyd and Taylor (1994) ecosystem respiration model (Eq.2) considering a dynamic temperature sensitivity (its temporal variability).

Reichstein et al. (2005) defined the constants values in Lloyd Taylor (1994) equation as follows; T_{ref} : 10°C and T_0 : -46.02°C. For E_0 , a constant value for each year was derived, while R_{ref} was estimated for every five days using a 15-day window. The difference between R_{eco} and measured NEE yields GPP estimate. In this respect, online CarboEurope partitioning tool (Reichstein et al., 2005) (available at <http://www.bgc-jena.mpg.de/~MDIwork/eddyproc/index.php>) was used. The filtered flux data (non-gap-filled) were put into the required format using R with R-Studio software program. The online tool used resulted in NEE and R_{eco} estimates, and then, GPP was calculated as the difference between NEE and R_{eco} . Negative or non-zero GPP values at nighttime are considered to be errors as no photosynthesis occurs at night (Thomas et al., 2011).

The performance of the model was tested applying a regression analysis for R_{eco} calculated by the model versus measured by the EC system (Table 4.7). Results for goodness of fits explained by adjusted coefficient of determination (r^2_{adj}) varied between 30.3 and 45.9%. Reichstein et al. (2005) compared the model performance to continuous soil chamber respiration data and reported r (correlation coefficient) values of 21% and 44% for two sites. Although the Reichstein's (2005) flux partitioning method have been used in the literature (e.g. Thomas et al., 2011; Lasslop et al., 2012; Morrison et al., 2013), the model performance has not been stated to the author's best knowledge.

Table 4.7 Regression analysis of R_{eco} values by model (Reischstein et al., 2005) versus $NEE_{nighttime}$ values by EC measurements

Year	T	$p <$	VIF	r^2	r_{adj}^2
2010	27.34	0.001	1.000	30.3	30.2
2011	37.23	0.001	1.000	42.5	42.5
2012	36.85	0.001	1.000	46.0	45.9
2013	4.73	0.001	1.000	28.8	28.7

T: t-statistics; VIF: variance inflation factor

Daily mean -GPP, R_{eco} and NEE fluxes from August of 2010 to mid-June of 2013 are illustrated in Figure 4.24. The model overpredicted daily -GPP peaks during August to September of 2010 (-0.25 to $-0.3 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$), May to July of 2011 and 2012 (-0.4 to $-0.45 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$), and April-May for 2013 (-0.3 to $-0.6 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$). The minimum -GPP daily fluxes (Table 4.8) were obtained during winter (December to March) of 2011, 2012 and 2013 (not measured for December). R_{eco} showed a decreasing trend during fall and winter and increased during spring and summer (Figure 4.24). R_{eco} reached higher peaks during August to September of 2010 than R_{eco} peaks during 2011 and 2012. R_{eco} flux was higher than GPP flux during summer in 2010 that caused positive NEE fluxes (CO_2 loss from the ecosystem to the atmosphere). The data also showed net CO_2 release during some summer days as a result of a sudden decrease in GPP flux.

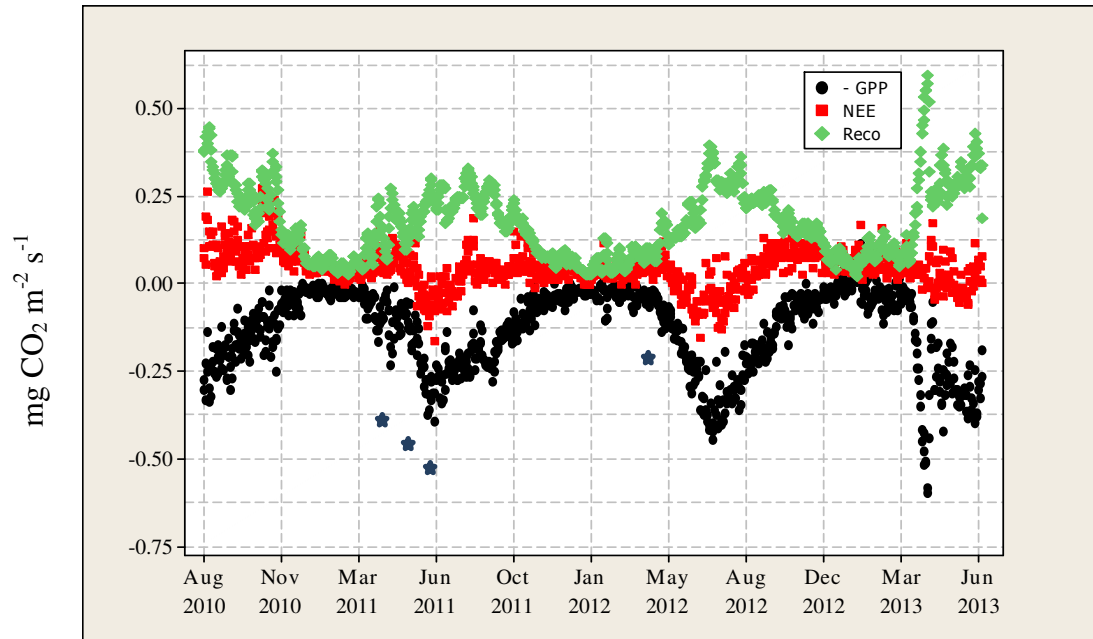


Figure 4.24. GPP, NEE and R_{eco} fluxes from August of 2010 to June of 2013. Points are one-day averages of measured data. ★ : Flooding period (April to June of 2011 and April of 2012).

Table 4.8. Descriptive statistics for NEE, R_{eco} and GPP ($\text{mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) fluxes for 2010, 2011, 2012 and 2013

Year	Variable	<i>N</i>	Mean	SD	CoefVar.	Min.	Median	Max.
2010	NEE	153	0.097	0.052	53.2	0.019	0.089	0.272
	R_{eco}		0.224	0.104	46.39	0.047	0.237	0.447
	GPP		0.127	0.0874	68.74	0.0007	0.123	0.34
2011	NEE	365	0.033	0.044	134.3	-0.167	0.034	0.215
	R_{eco}		0.148	0.083	56.14	0.026	0.149	0.331
	GPP		0.118	0.094	79.60	0.0009	0.095	0.4
2012	NEE	333	0.031	0.056	177.9	-0.154	0.037	0.167
	R_{eco}		0.164	0.089	54.52	0.024	0.155	0.395
	GPP		0.135	0.119	88.64	0.0004	0.087	0.45
2013	NEE	170	0.03	0.039	131.9	-0.059	0.036	0.17
	R_{eco}		0.22	0.134	66.11	0.046	0.24	0.594
	GPP		0.192	0.15	77.8	0.60	0.183	0.01

*mean, standard deviation (SD), coefficient of variance (CoefVar.), minimum (Min.), median and maximum (Max.) values of NEE, R_{eco} and (-)GPP fluxes for measured days (*N*) in 2010, 2011, 2012 and 2013 years are presented in the table. Negative indicates uptake by the ecosystem (loss from the atmosphere)

The seasonal components of NEE, GPP and R_{eco} fluxes are introduced in Table 4.9. The data were available from August to December for 2010, thus the summer was calculated with only August and September and the winter only included December. In addition, the measurement period was ended in the mid-June of 2013, so the only complete data were available for spring of 2013. A reliable comparison of seasonal GPP and R_{eco} components was performed using the complete period data. The highest GPP and the lowest R_{eco} fluxes were observed in summer and winter, respectively, both in 2011 and 2012. The spring R_{eco} flux was highest in 2013 followed by 2011 and 2012. The situation was quite an opposite for GPP flux which resulted in a lower NEE in 2013. The highest R_{eco} and GPP fluxes were in 2010 in the fall. R_{eco} flux was higher in 2012 than in 2011, also GPP flux was lower in 2012 than in 2011, which caused higher source strength for 2012.

4.3.7. Modeling of NEE, R_{eco} and GPP

The correlations between environmental factors (T_{air} , VPD, T_{soil} , ET and RF) and the carbon components (NEE, R_{eco} and GPP) were investigated using a best subset procedure to build linear regression models. The correlation coefficient matrix was performed over hourly data for each year (2010, 2011, 2012 and 2013) which are introduced in Tables 4.10 to 4.13. The carbon components had highest correlations with T_{air} , VPD and ET for all the years. The lowest correlations were obtained between carbon components and RF for 2011, 2012 and 2013. There were no significant relationships between GPP, R_{eco} and RF for 2010 ($p > 0.01$). The strongest relationships were obtained between NEE and VPD ($0.44 < r < 0.48$); R_{eco} and T_{air} ($0.71 < r < 0.88$); and GPP and ET ($0.71 < r < 0.82$).

Table 4.9 . Seasonal components of observed Net Ecosystem Exchange of CO₂ (NEE) and modeled Sums of Photosynthesis (GPP) and Respiration (R_{eco})

Period	2010			2011			2012			2013		
	NEE	R_{eco}	GPP	NEE	R_{eco}	GPP	NEE	R_{eco}	GPP	NEE	R_{eco}	GPP
Spring	NA	NA	NA	252.4	892.5	640.1	105.9	755.8	649.9	80.79	1713.9	1633.1
Summer	519.7	1607.1	1087.4	173.9	2431.3	2257.4	56.3	2652.3	2596	18.5	330.9	312.4
Fall	607.7	1103.4	495.7	226.1	534.9	308.8	434.8	712.6	277.8	NA	NA	NA
Winter	165.7	230.3	64.6	387.9	129.8	258.1	302.7	484.9	182.2	423.4	752.8	329.4
Annual	1293.1	2940.8	1647.7	1040.3	4504.7	3464.4	899.7	4605.6	3705.9	522.7	2797.6	2274.9
g C m ⁻² period ⁻¹	352.7	802.04	449.4	283.7	1228.6	944.9	245.4	1256.1	1010.7	142.6	762.9	620.3

*Annual period is from 1 August to 31 December for 2010, from 1 January to 31 December for 2011 and from 1 January to 28 November for 2012. Seasonal periods are Spring (April-May), Summer (June-July-August-September), Fall (October-November), and Winter (December-January- February-March). All values are in g CO₂ m⁻² period⁻¹ unless otherwise stated. NA: not measured.

Correlation analyses showed that NEE had stronger relationships with GPP ($0.81 < r < 0.93$) when compared to R_{eco} ($0.14 < r < 0.27$). Also, results indicated that the relationships between R_{eco} and GPP was moderate ($0.56 < r < 0.69$).

Table 4.10. Correlation matrix (r) of average hourly NEE and its components (GPP and R_{eco}) and ancillary data for 2010

	NEE	T _{air}	VPD	T _{soil}	ET	RF	GPP
T _{air}	-0.291						
VPD	-0.442	0.817					
T _{soil}	-0.247	0.996	0.797				
ET	-0.695	0.734	0.858	0.706			
RF	0.065	-0.039	-0.087	-0.03*	-0.062		
GPP	-0.809	0.661	0.632	0.619	0.816	-0.03*	
R_{eco}	-0.141	0.758	0.525	0.738	0.526	0.04*	0.697

Correlation is significant at the $\alpha=0.01$ level, * p value is smaller than 0.01.

Table 4.11. Correlation matrix (r) of average hourly NEE and its components (GPP and R_{eco}) and ancillary data for 2011

	NEE	T _{air}	VPD	T _{soil}	ET	RF	GPP
T _{air}	-0.274						
VPD	-0.450	0.729					
T _{soil}	-0.256	0.995	0.699				
ET	-0.664	0.650	0.837	0.612			
RF	0.018	0.023	-0.056	0.024	-0.035		
GPP	-0.906	0.605	0.630	0.573	0.784	-0.007	
R_{eco}	-0.160	0.875	0.591	0.882	0.538	0.019	0.563

Correlation is significant at the $\alpha=0.05$ level

Table 4.12. Correlation matrix (r) of average hourly NEE and its components (GPP and R_{eco}) and ancillary data for 2012

	NEE	T _{air}	VPD	T _{soil}	ET	RF	GPP
T _{air}	-0.313						
VPD	-0.457	0.770					
T _{soil}	-0.056	0.811	0.470				
ET	-0.706	0.649	0.827	0.327			
RF	-0.012	-0.065	-0.058	-0.092	-0.026		
GPP	-0.927	0.585	0.627	0.374	0.796	-0.02	
R_{eco}	-0.267	0.841	0.644	0.842	0.551	0.08	0.609

Correlation is significant at the $\alpha=0.05$ level

Table 4.13. Correlation matrix (r) of average hourly NEE and its components (GPP and R_{eco}) and ancillary data for 2013

	NEE	T_{air}	VPD	T_{soil}	ET	RF	GPP
T_{air}	-0.358						
VPD	-0.476	0.799					
T_{soil}	-0.069	0.714	0.381				
ET	-0.602	0.691	0.841	0.313			
RF	0.041	-0.034	-0.084	-0.006	-0.08		
GPP	-0.871	0.630	0.632	0.421	0.712	-0.03	
R_{eco}	-0.266	0.719	0.553	0.727	0.527	-0.01	0.705

Correlation is significant at the $\alpha=0.05$ level

Environmental variables selected were ET, T_{air} and VPD as explanatory variables owing to their high correlation with the carbon components. Thus, regression analyses were performed using the best subset procedure as a function of these explanatory variables using Minitab 16. Results of the best subset regressions are presented in Tables 4.14 to 4.17 for 2010 to 2013, respectively. The p -value and variance inflation factor (VIF) indicated that the relationship between the predictors (ET, T_{air} and VPD) and the responses (NEE, GPP and R_{eco}) are statistically significant ($p < 0.05$) and also there was no multicollinearity ($VIF < 10$). The regression models obtained for GPP and R_{eco} showed good r^2_{adj} values ($52\% < r^2_{adj} < 78\%$) when compared to NEE ($36\% < r^2_{adj} < 61\%$). Strongly and weakly predictive models for NEE, GPP and R_{eco} were observed for 2010 and 2013, respectively. This can be explained by the lack of complete measurements in these years. The measurement periods mostly included summer and fall months for 2010 but mostly included winter and spring months for 2013. Due to the unavailable meteorological conditions, the missing data were higher during winter and spring than during summer and fall that may have caused a higher percentage of gap-filled data percentage in 2013 than in 2010 (Table 4.3) and may have adversely affected the model quality.

Table 4.14. Best regression models of NEE, GPP and R_{eco} as a function of environmental variables for 2010

	Explanatory variables					Model summary				
Response	Predic.	Coeffs	T	<i>P</i>	VIF	r^2	r^2_{adj}	F	<i>P</i>	<i>n</i>
NEE	Const.	0.032	8.9	0.001		61	61	1395	0.001	2676
	ET	-0.951	-52.4	0.001	3.83					
	T _{air}	0.006	15.84	0.001	3.05					
	VPD	0.006	12.46	0.001	5.32					
GPP	Const.	0.046	11.2	0.001		72.1	72.1	2306	0.001	2676
	ET	1.05	49.8	0.001	3.83					
	T _{air}	0.008	18.9	0.001	3.05					
	VPD	-0.012	-21.1	0.001	5.32					
R_{eco}	Const.	0.078	26.71	0.001		60.8	60.8	1384	0.001	2676
	ET	0.104	6.95	0.001	3.83					
	T _{air}	0.014	46.11	0.001	3.05					
	VPD	-0.006	-14.8	0.001	5.32					

Predic.: predictors; Coeffs.: coefficients; T:t-test statistics; VIF: variance inflation factor; r^2 :coefficient of determination; r^2_{adj} : adjusted coefficient of determination; *n*: sample size; F: F-test statistics

Table 4.15. Best regression models of NEE, GPP and R_{eco} as a function of environmental variables for 2011.

	Explanatory variables					Model summary				
Response	Predic.	Coeffs	T	<i>P</i>	VIF	r^2	r^2_{adj}	F	<i>P</i>	<i>n</i>
NEE	Const.	0.071	38.7	0.001		49.5	49.4	2793	0.001	8567
	ET	-1.05	-69.6	0.001	3.4					
	T _{air}	0.004	16.9	0.001	2.2					
	VPD	0.001	14.7	0.001	4.2					
GPP	Const.	0.004	2.1	0.001		64.5	64.4	5178	0.001	8567
	ET	1.05	69.94	0.001	3.4					
	T _{air}	0.006	25.51	0.001	2.2					
	VPD	-0.01	-18.16	0.001	4.2					
R_{eco}	Const.	0.07	122.5	0.001		77.9	77.9	1082	0.001	8567
	ET	0.005	0.99	0.001	3.4					
	T _{air}	0.01	127.5	0.001	2.2					
	VPD	-0.002	-10.5	0.001	4.2					

Predic.: predictors; Coeffs.: coefficients; T:t-test statistics; VIF: variance inflation factor; r^2 :coefficient of determination; r^2_{adj} : adjusted coefficient of determination; *n*: sample size; F: F-test statistics

Table 4.16. Best regression models of NEE, GPP and R_{eco} as a function of environmental variables for 2012.

Response	Explanatory variables					Model summary				
	Predic.	Coeffs	T	P	VIF	r^2	r^2_{adj}	F	P	n
NEE	Const.	0.084	41.1	0.001		55.6	55.6	3487	0.001	8360
	ET	-1.194	-80.35	0.001	3.2					
	T _{air}	0.002	10.41	0.001	2.5					
	VPD	0.01	20.2	0.001	4.6					
GPP	Constc	-0.003	-1.23	0.001		65.8	65.8	5355	0.001	8360
	ET	1.21	76.4	0.001	3.2					
	T _{air}	0.006	23.05	0.001	2.5					
	VPD	-0.01	-19.93	0.001	4.6					
R_{eco}	Const.	0.082	103.7	0.001	3.2	70.7	70.7	6734	0.001	8360
	ET	0.017	2.93	0.001	2.5					
	T _{air}	0.008	91.2	0.001	4.5					
	VPD	-0.001	-2.6	0.001						

Predic.: predictors; Coeffs.: coefficients; T:t-test statistics; VIF: variance inflation factor; r^2 :coefficient of determination; r^2_{adj} : adjusted coefficient of determination; n : sample size; F: F-test statistics

Table 4.17. Best regression models of NEE, GPP and R_{eco} as a function of environmental variables for 2013.

Response	Explanatory variables					Model summary				
	Predic.	Coeffs	T	P	VIF	r^2	r^2_{adj}	F	P	n
NEE	Const.	0.103	26.56	0.001		36.9	36.9	789.9	0.001	4045
	ET	-0.93	-30.35	0.001	3.4					
	T _{air}	0.003	4.63	0.001	2.8					
	VPD	0.002	1.40	0.001	4.9					
GPP	Constc	0.003	0.70	0.001		54.6	54.5	1619	0.001	4045
	ET	1.08	30.62	0.001	3.4					
	T _{air}	0.012	17.85	0.001	2.8					
	VPD	-0.007	-5.37	0.001	4.9					
R_{eco}	Const.	0.106	45.52	0.001		52.5	52.4	1489	0.001	4045
	ET	0.153	8.30	0.001	3.4					
	T _{air}	0.015	41.94	0.001	2.8					
	VPD	-0.006	-7.98	0.001	4.9					

Predic.: predictors; Coeffs.: coefficients; T:t-test statistics; VIF: variance inflation factor; r^2 :coefficient of determination; r^2_{adj} : adjusted coefficient of determination; n : sample size; F: F-test statistics

4.4. Water and Energy fluxes

4.4.1. Water fluxes

Rainfall (RF) and evapotranspiration (ET) are the influx and outflux components of hydrological cycle for water budget (WB) and were continuously measured between 01 August 2010 and 16 June 2013. ET increased during the growing seasons (summer and spring) and decreases the non-growing seasons (winter and fall) (Figure 4.25). The maximum monthly mean ET occurred in August of 2010 ($0.2 \text{ mm month}^{-1}$), July of 2011 ($0.19 \text{ mm month}^{-1}$), July of 2012 ($0.2 \text{ mm month}^{-1}$) and May of 2013 ($0.18 \text{ mm month}^{-1}$). The minimum monthly mean ET was obtained in December of 2010 ($0.4 \text{ mm month}^{-1}$), January of 2011, 2012 and 2013 ($< 0.3 \text{ mm month}^{-1}$).

The WB was defined according to the cumulative sums of ET (negative flux) and PPT (positive flux) ignoring stream discharges (Section 2.3) for each year (i.e. $\text{WB} \approx \text{PPT} - \text{ET}$). The WB was generally negative because of higher cumulative ET (Figure 4.26) during the measurement period. 2012 was the most arid year and had higher ET that indicated water deficit for the peatland. ET and RF were generally similar during January to March in 2012 and 2013; however, this period was extended until June in 2011.

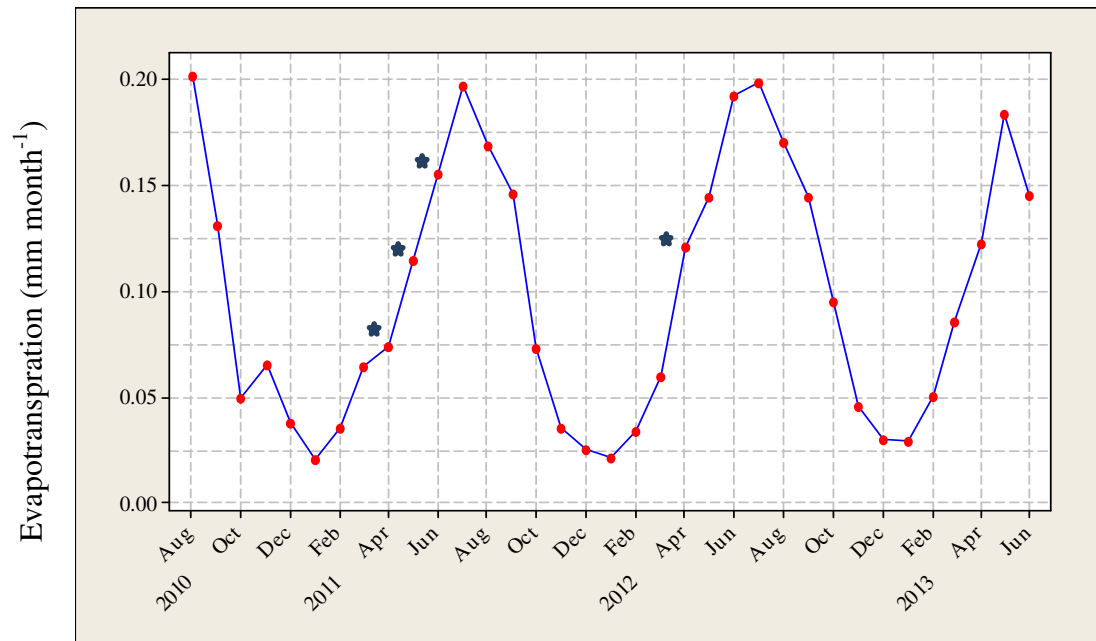


Figure 4.25. Monthly mean ET for the period of 01 August 2010 to 19 June 2013. ★ : Flooding period.

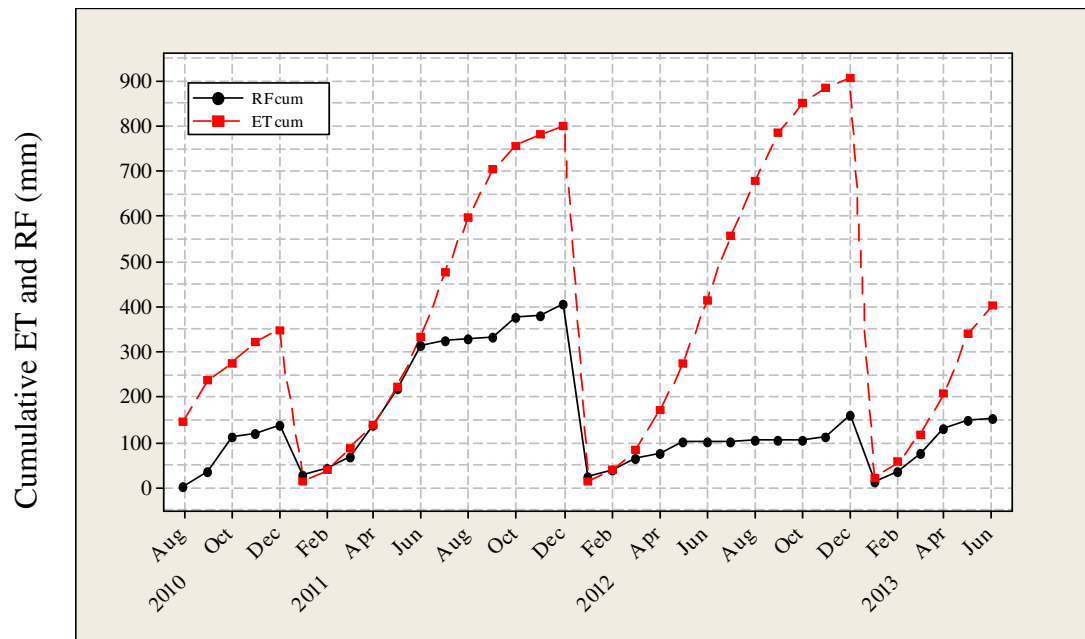


Figure 4.26. Cumulative sums of ET and RF between 01 August 2010 and 19 June 2013.

4.4.2. Energy fluxes

Energy balance closure is used to assess the performance of EC flux system (Lund et al., 2007). According to the energy balance Eq. (2.9) (Section 2.3), net radiation (R_n) is partitioned into the latent heat (LE), sensible heat (H) and ground heat fluxes (G) (Burba et al., 1999). H is used for warming the air; LE is used for water evaporation and plant transpiration (i.e. evapotranspiration); and G is used to heat the soil. Under a perfect closure, the sum of the sensible and latent heat flux (H+LE) measured by EC is equal to the difference between net radiation and ground (soil) heat flux ($R_n - G$) measured independently from the meteorological sensors (Chapter 3) (Aubinet et al., 2012).

In this study, soil heat flux (G) could not be measured because of the lack of available instrument for this measurement. Direct measurements of G can be made using a heat flux plate whose position should be stable under a few centimeters below the soil surface (Ronda et al., 2008). However, water flow can disturb the heat plates thus hindering G measurements. Thus, Wang et al. (1999) improved a sinusoidal analytical method and a half-order time derivative method for the long-term estimation of G. Hsieh et al. (2008) improved this analytical method achieving a good agreement between the measured and estimated G values ($r^2 > 0.85$). This improved method is based on the following Eq. (4.4);

$$G = 2 \sqrt{\frac{k \rho_s c_s}{\pi}} \sum_{i=0}^N \frac{T_{i+1} - T_i}{t_{i+1} - t_i} [\sqrt{t_N - t_{i+1}} - \sqrt{t_N - t_i}] \quad (4.4)$$

where k: thermal conductivity ($W m^{-1} K^{-1}$); ρ_s : soil density ($kg m^{-3}$); c_s : soil

specific heat ($\text{J kg}^{-1}\text{°K}^{-1}$); while $\rho_s c_s$ ($\text{J m}^{-3}\text{°K}^{-1}$) represents the volumetric heat capacity; T : soil temperature (°K); and t : time.

In this study, the volumetric heat capacity and the thermal conductivity of the peat could not be determined. Therefore, the related literature values for the peat soil were used instead. The following values are used for the G calculation:

$$\rho_s c_s = 45 \times 10^6 \text{ J m}^{-3}\text{°K}^{-1} \text{ (Tsuang, 2005); } k = 0.71 \text{ W m}^{-1}\text{°C}^{-1} \text{ (Hanson et al., 2000).}$$

The soil temperature (T) was recorded each second for each year during the measurement period. Thus, G was calculated according to the Eq. (4.4) by substituting the $\rho_s c_s$ and k values for peat soil using Minitab 16. Once the G was calculated for each year, the energy balance closure was estimated. $H+LE$ was plotted as a function of R_n-G (Figure 4.27-4.30) and the regression analysis was performed using Minitab 16 (Table 4.18). The relationships between R_n-G and $H+LE$ were reported with their slopes of 0.94, 0.82, 0.81 and 0.88 for 2010, 2011, 2012 and 2013, respectively.

The coefficients of R_n-G explained the amount of net energy used as latent and sensible energy (i.e. available energy) for each year. The percentages of the energy used for the evapotranspiration and warming the air were 71%, 62%, 64% and 51% for 2010, 2011, 2012 and 2013, respectively (Table 4.18).

Table 4.18. Best regression models of H+LE as a function of R_n-G for 2010, 2011, 2012 and 2013

	Response	Explanatory					Model summary			
		Predic.	Coeffs	T	P	VIF	r^2	F	P	n
2010	H+LE	Const.	-2.44	-3.1	0.002		94.4	52195	0.001	3094
		R_n-G	0.71	228.5	0.001	1.000				
2011	H+LE	Const.	11.2	9.8	0.001		82.4	25317	0.001	5408
		R_n-G	0.62	159.1	0.001	1.000				
2012	H+LE	Const.	1.8	1.3	0.002		81	20651	0.001	4831
		R_n-G	0.64	143.7	0.001	1.000				
2013	H+LE	Const.	-14.9	-16.4	0.001		88.3	30785	0.001	4068
		R_n-G	0.51	175.5	0.001	1.000				

Predic.: predictors; Coeffs.: coefficients; T:t-test statistics; VIF: variance inflation factor; r^2 :coefficient of determination; n:measurement number; F: F-test statistics

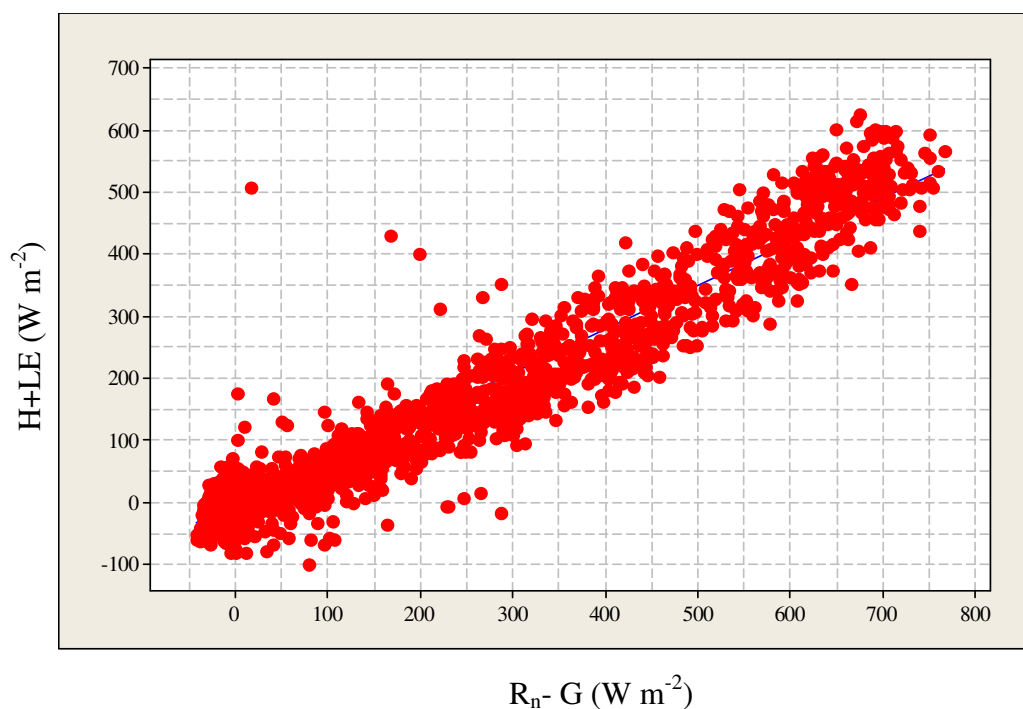


Figure 4.27. Energy balance closure for 2010.

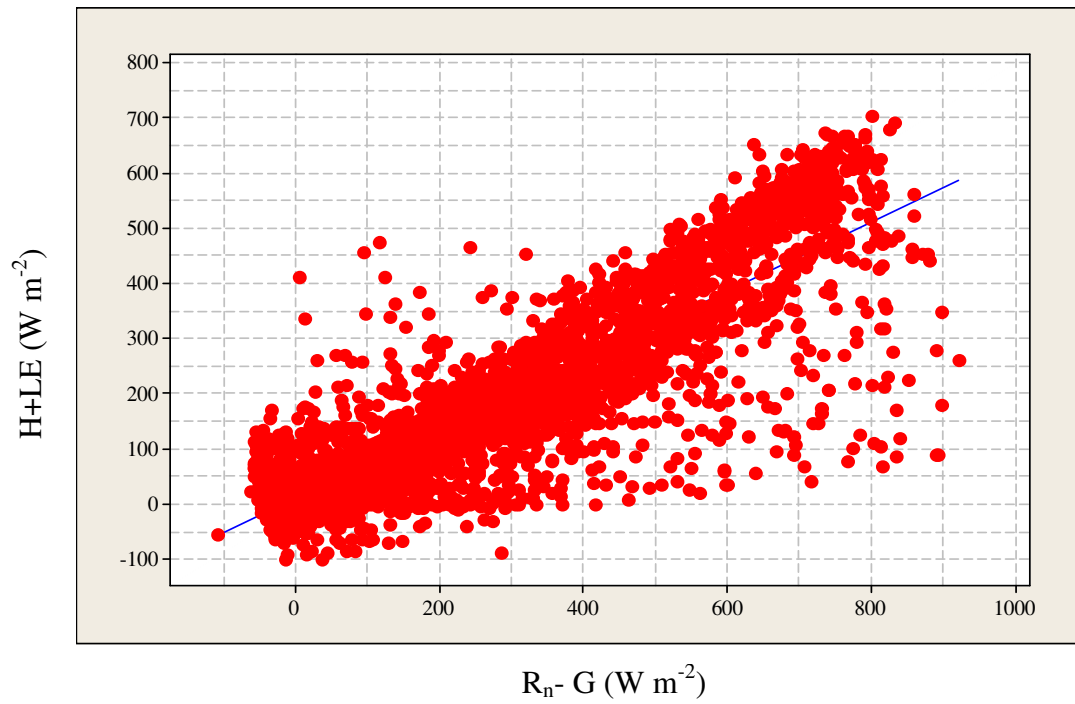


Figure 4.28. Energy balance closure for 2011.

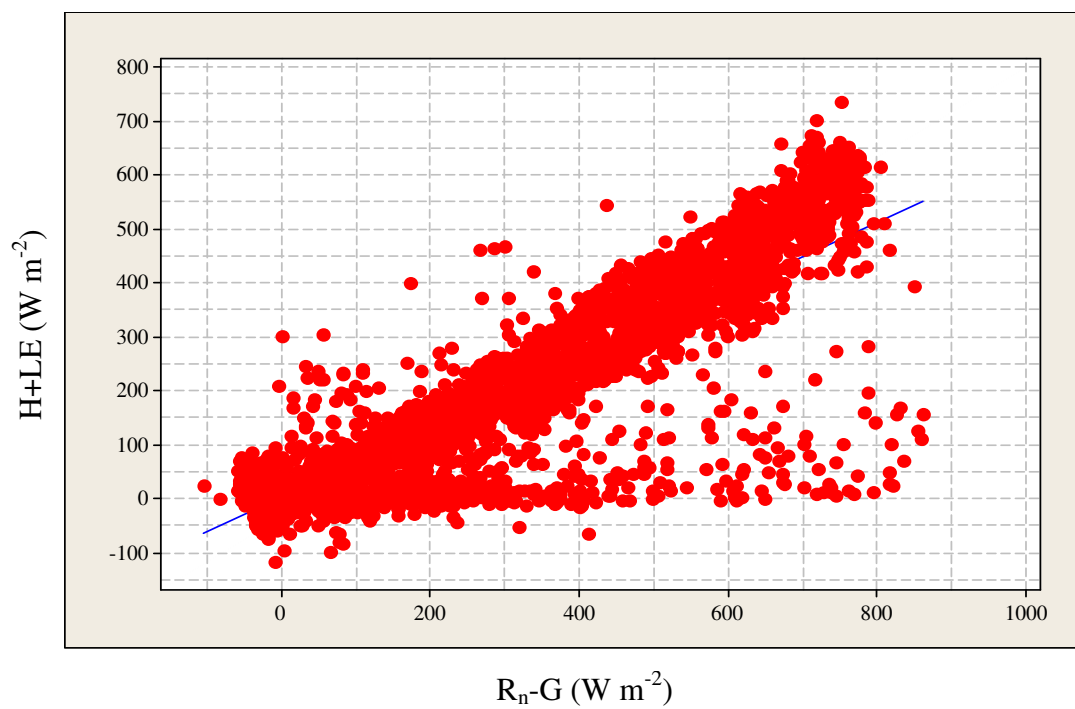


Figure 4.29. Energy balance closure for 2012.

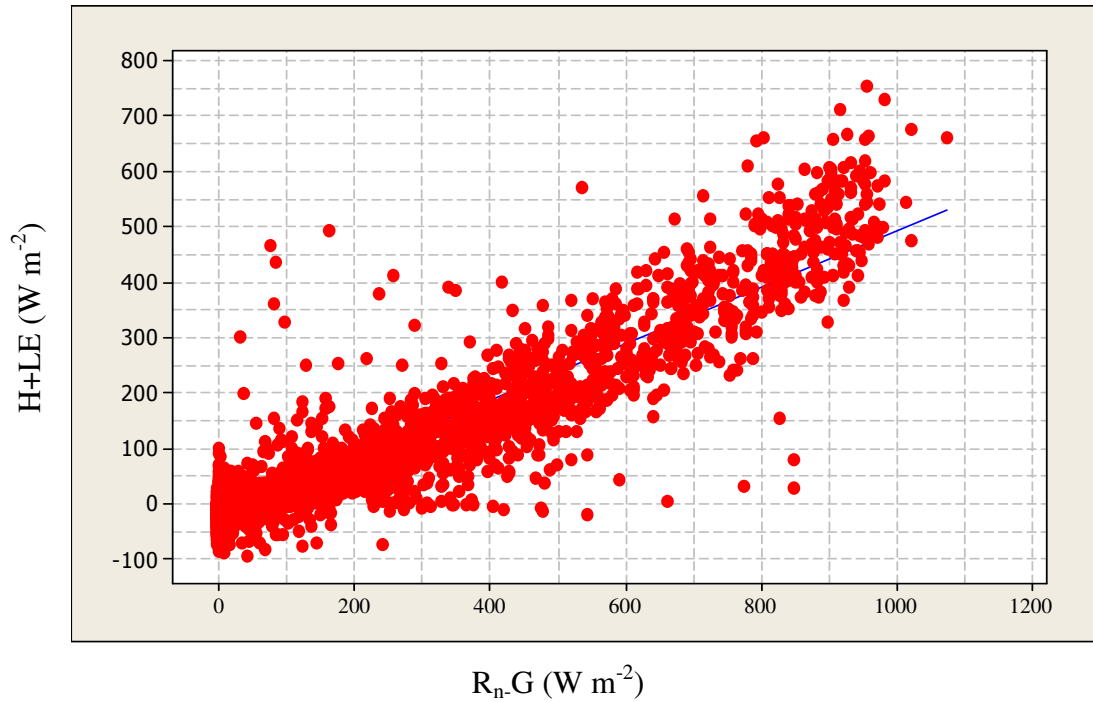


Figure 4.30. Energy balance closure for 2013.

Bowen ratio (β) gives an indication of relative importance of latent heat flux to sensible heat flux, and thus, can be used as an aridity index of an ecosystem (Arora, 2002; Vaccari et al., 2004; Gerrits et al., 2009; Dunkel, 2009). Monthly β ratios were estimated according to the Eq. (2.14) for each year (Figure 4.31). Negative values were obtained for January to March of 2012 and December of 2010 and 2011. β gets negative when H is low, around sunrise, sunset, or occasionally at night (Erell et al., 2011). This situation occurs more often with cold weather (Erell et al., 2011). The magnitudes of β showed a similar trend for each year in that higher β values (0.3-1.1) were obtained during the growing season (April to September) extending to the fall season (October to November), and lower β values (-1.5 to 0.2) were obtained during the winter season (December to March). The β ratio values of the site showed the characterization of a humid climate during the measurement period.

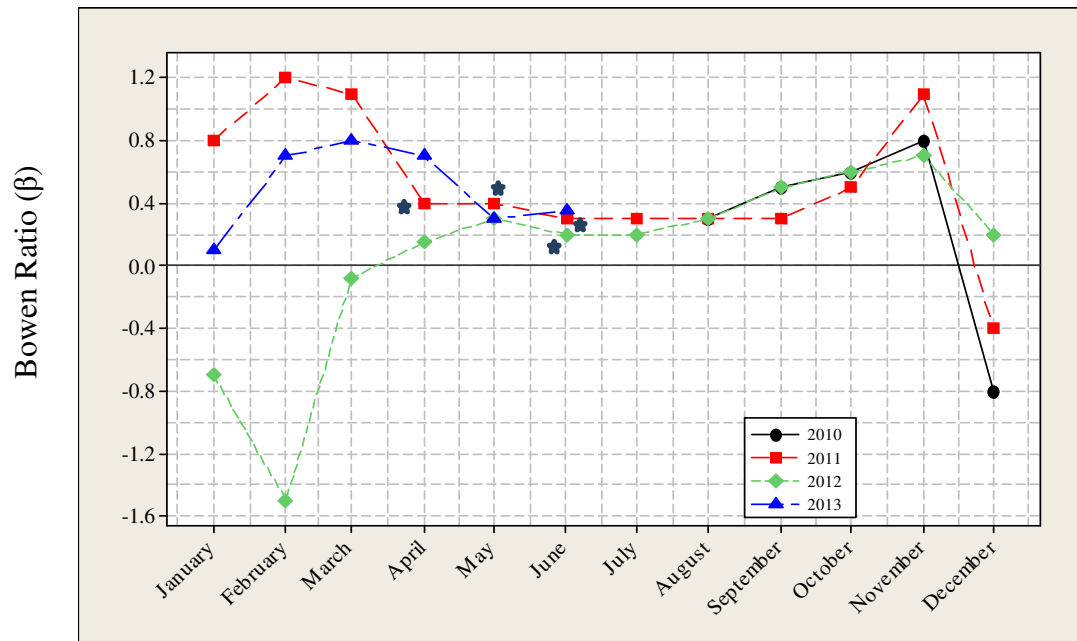


Figure 4.31. Monthly variation of Bowen ratio (β) in 2010, 2011, 2012 and 2013.

★: Flooding period.

CHAPTER IV

5. DISCUSSIONS

5.1. Annual CO₂ exchange

CO₂ source strength of Yeniçağa peatland was higher during the non-growing season and lower during the growing season, and Yeniçağa peatland switched to sink in June of 2011 and May to July of 2012 (Figure 4.22). The cumulative annual NEE was estimated at 283.8 g C m⁻² in 2011 and 252.4 g C m⁻² in 2012 (Table 4.6) which showed that the peatland acted as a strong source of CO₂. The source strength of Yeniçağa peatland was at least 2 times higher than the peatlands of a similar climate regime (Table 5.1). The comparable studies showed that strong CO₂ uptake during the growing season led to the peatland acting as a strong sink of CO₂ when no extreme weather conditions occurred (e.g. Strilesky et al., 2012). Only with extreme meteorological conditions (e.g. drought, high rainfall, and flooding) the sink strength decreased (Lund et al., 2012) and turned into a source of CO₂ (Nieveen et al., 1998; Shurpali et al., 1995). Flooding events observed for Yeniçağa peatland revealed a decline in NEE in this study, as was also observed in the related literature. The flooded events occurred during April to June of 2011 and April of 2012, and the source strength of CO₂ increased in April of 2011 and 2012 even though the month

of April marked the beginning of the growing season for Yeniçağa peatland. Then, the source strength of the peatland decreased in May of 2011 and turned into a sink of CO₂ in June of 2011. However, the situation for May and June of 2012 and July of 2012 was different in that the peatland switched to a sink of CO₂. As a result, the flooding initially increased the peatland source strength but its impact on the source strength decreased in the following months. This result also appears to be supported by discussion afterwards about effect of high rainfall event on CO₂ fluxes in the following paragraph.

Some studies reported that higher rainfall events caused an increase in CO₂ fluxes to the atmosphere (Kim et al., 2012; Munson et al., 2010; Lee et al., 2002). One possible explanation is that increased rainfall could lead to increases in soil microbial activity and soil respiration (Franzluebbers et al., 2000; Harper et al., 2005). Conversely, some studies reported that soil CO₂ flux was suppressed by an increase in precipitation events (Lee et al., 2002). The decrease in CO₂ flux after heavy precipitation might have resulted from reduced air-filled soil pores, and thus, from decreased soil respiration with increased anaerobism (Lee et al., 2002). In addition, discussions on the effects of precipitation on CO₂ flux showed that the variability of ecosystem CO₂ flux depends on the duration, severity, timing and pulse of precipitation event, and the aridity conditions of the site prior to precipitation (Fan et al., 2012; Wu et al., 2011; Munson et al., 2010; Patrick et al., 2007; Huxman et al., 2004).

A better understanding of ecosystem responses to increased rainfall events necessitates long-term measurement and monitoring. However, in this study, the

effects of flooding events showed that flooding had a positive effect on source strength initially which means a higher ecosystem respiration (soil and plant respiration) for which increased soil respiration is most likely to be responsible. In the following period of the flooding, net primary productivity (NPP) became more dominant than soil respiration which resulted in a decrease in the source strength of the peatland (Figure 4.22). To better understand this relationship, variation in SWC with NEE was also considered. A sharp increase in SWC in April of 2011 increased NEE. NEE decreased in May to June of 2011 when the rate of increase in SWC declined (Figure 5.1). This case was similar to findings reported by Kim et al. (2012). When the same period was considered for 2012, an increase was observed in both SWC and NEE in April of 2012.

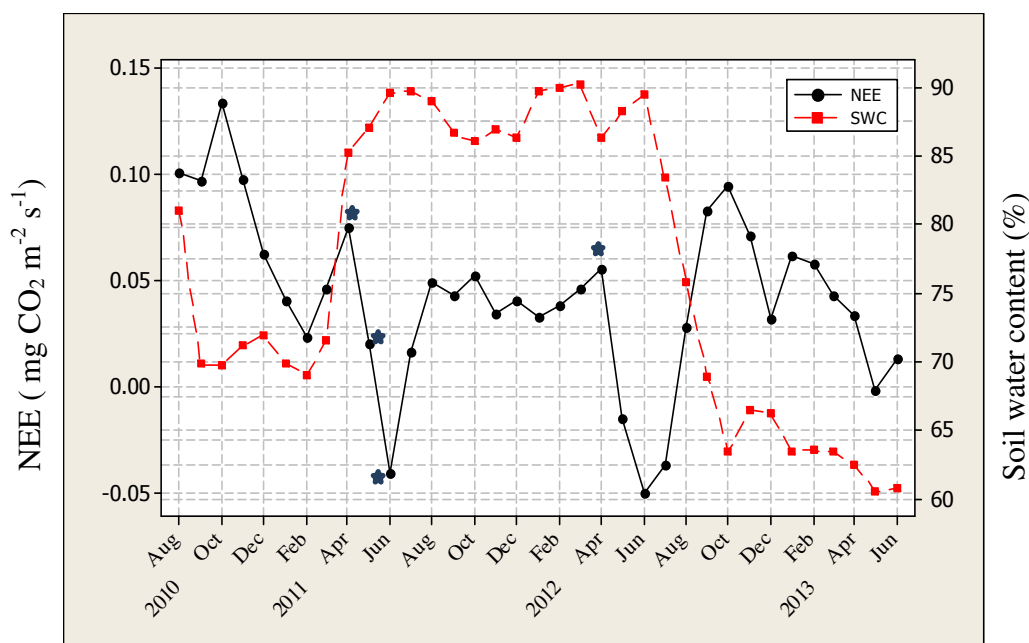


Figure 5.1. Monthly variations in NEE and SWC from August 2010 to June 2013.

★ : Flooding period.

Since SWC was already high before the flooding period in 2012, the increases in SWC and NEE were of smaller magnitudes than those of April of 2011, and this did not lead to a significant change in NEE. The cumulative NEE in 2011 was higher than in 2012 (a higher source strength in 2011). This result can be related to the effects of the flooding period on SWC. The long flooding period (approximately 3 months) happened in 2011 that caused an increase in SWC and maintained higher SWC levels in the following months. The site was flooded only for one month in 2012 which did not lead to as much increase in SWC in the following months of 2012 as in 2011.

The drought period should also be discussed in relation to changes in NEE. In this study, the lowest precipitation events during August to October of 2012 (Figure 4.1) resulted in a sharp decrease in SWC ($< 70\%$, Figure 5.1). During that period, there was a sharp increase in the source strength of CO_2 (Figure 5.1). In addition, a similar pattern (SWC $< 73\%$ and increased source strength) was observed during the fall season of 2010. The drought period resulted in a higher source strength for 2010 and 2012 than for 2011 (Table 4.9). Given the GPP and R_{eco} values, the drought period mostly affected R_{eco} relative to GPP (Table 4.9). The ecosystem response to the drought period is still under discussion (Lund et al., 2012) as with higher precipitation events. Some studies suggested that GPP was affected primarily (Shurpali et al., 1995; Arneth et al., 2002; Lafleur et al., 2003), while others argued that R_{eco} is more affected (Aurela et al., 2007) in peatlands, under drought conditions. In addition, Lund et al. (2012) indicated that the timing, severity and duration of a drought control drought effects on GPP, R_{eco} and NEE of a temperate peatland.

Although the flooding and drought periods may have increased the CO₂ source strength of the peatland, these factors cannot be held as only explanations for the high source strength relative to the related literature (Table 5.1). Changes in land use and land cover (LULC) can be considered to contribute a great deal to the peatland acting as a strong source of CO₂. Drainage, the use of peatlands for agriculture, overgrazing, and peat mining appear to enhance the source strength of the Yeniçağa peatland. Quantification of LULC-induced disturbances of Yeniçağa peatland was dated as far back as 1944 based on remotely-sensed data and change-detection techniques (Evrendilek et al., 2011). The site has been used heavily for overgrazing. In fact, some studies reported high positive NEE magnitudes (i.e. a high source strength of CO₂) for disturbed peatlands (Table 5.2). For example, Lohila et al. (2008) reported an annual NEE of 210 g C m⁻² yr⁻¹ for a cultivated peatland in southern Finland which is similar to annual NEE results of this study for 2011 and 2012. Similarly, NEE measurements were made in two locations for a peatland in Quebec (Canada) where peat extraction was continued for eight years (1972 to 1980) and then restoration activities took place to compare restored versus unrestored area (Strack et al., 2013). Results showed that the unrestored peatland was a significant source of CO₂ (with an annual NEE of 343.5 g C m⁻²) as with this study. The restoration activities caused a decrease in the source strength of the peatland. Also, Morrison et al. (2013) reported a NEE value of 270 g C m⁻² from July to October (120 days) for a drained and intensively cultivated peatland even with the measurement period of non-growing season not included, which is in agreement with the annual NEE of this study (Table 5.2).

Table 5.1. Annual NEE ($\text{g C m}^{-2} \text{ yr}^{-1}$) for Yeniçağa peatland and peatlands with similar climate regime for a comparison

Site	Location	Climate zone	Latitude	NEE	Time Period	References
Mer Bleue	Canada	Cool-temperate	45 24 N 75 29 W	-104	2006-2007	Strilesky et al., 2012
Fajemyr	Southern Sweden	Temperate	56 15 N 13 33 E	-21.4 14.3 -29.4 23.6 -28.9	2005 2006 2007 2008 2009	Lund et al., 2007 Lund et al., 2012
Glencar	Southwest Ireland	Temperate	51 55 N 9 55 W	-66.8 -67.2 -84.0 -12.5 -13.5 -42.7	2003 2004 2005 2006 2007 2008	Koehler et al., 2011
Panjin	China	Warm-temperate	121 54 E 41 08 N	-65	2005-2006	Zhou et al., 2009
Degero Stormyr	Northern Sweden	Cold-temperate	64 12 N 19 34 E	-55 -48	2004 2005	Nilson et al., 2008
Mer Bleue	Southeast Canada	Cool-temperate	45 41N 75 48 W	-40.2	1998-2004	Roulet et al., 2007
Fochteloer	North Netherland	Temperate	53 30 N 6 23 W	97	1994-1995	Nieveen et al., 1998
Chippewa	North central Minnesota	Temperate	47 32 N 93 28W	71 g C m^{-2} -32 g C m^{-2}	mid-May - mid-October 1991 mid-May - mid-October 1992	Shurpali et al., 1995
Yenicaga Lake	Northwest Turkey	Cool-temperate	40 47 N 32 1 E	352.7 283.8 252.4 142.6	2010(Aug-Dec) 2011 2012 2013 (Jan-Jun)	This study

Table 5.2. Annual NEE ($\text{g C m}^{-2} \text{ yr}^{-1}$) for Yeniçağa peatland and disturbed peatlands for a comparison

Site	NEE ($\text{g C m}^{-2} \text{ yr}^{-1}$)	Reference
Finland-Cultivated peatland	210	Lohila et al. (2008)
Canada-Peat extracted for 8 years	343.5	Strack et al. (2013)
England-drained and intensively cultivated	270 g C m^{-2} (Jul to Oct)	Morrison et al. (2013)
Yeniçağa peatland	352.7 g C m^{-2} (Aug-Dec) 283.8 252.4 142.6 g C m^{-2} (Jan-Jun)	This study

5.2. Relationship between carbon components

One of the necessities of the partitioning of NEE into GPP and R_{eco} is to improve the understanding of causes of inter- and intra-annual variability of NEE (Vickers et al., 2009). There was a stronger correlation (Table 4.10 – 4.13) between NEE and GPP ($0.81 < r < 0.85$) than between NEE and R_{eco} ($0.14 < r < 0.27$) which provided that variability in GPP primarily controls NEE. Although Valentini et al. (2000) suggested that variability in R_{eco} primarily controls the carbon balance, some studies showed that GPP may play a more important role than R_{eco} in determining variability in NEE (Reichstein et al., 2007; Stoy et al., 2008), as was supported by this study.

The relationship between GPP and R_{eco} is important to the predictions of impacts of climate change on future C sequestration by ecosystems (Vickers et al., 2009). The correlation between GPP and R_{eco} investigated in the related literature revealed a strong correlation ($0.63 < r < 0.94$) (e.g. Baldocchi, 2008; Stoy et al, 2008; Recihstein et al., 2007; Gilmanov et al., 2007). The correlation was analyzed between GPP and R_{eco} for each year in this study (Table 4.11-4.14) and results

showed a strong correlation between GPP and R_{eco} ($0.56 < r < 0.71$) which was consistent with the literature. However, the above studies and this study did not account for a spurious correlation that may arise since GPP and R_{eco} were derived from the same dataset (since $GPP = R_{eco} - NEE$) (Vickers et al., 2009). Only Lasslop et al. (2010) aimed to minimize the spurious correlation in reporting the correlation between GPP and R_{eco} ($r = 0.85$) derived from nighttime and daytime datasets.

The self-correlation was not taken into account in this study. In agreement with findings by Reichstein et al. (2007) and Stoy et al. (2008), the correlation coefficient matrix of this study appeared to show that changes in GPP had a primary control on NEE, while R_{eco} followed GPP variability.

5.3. Environmental controls on NEE, R_{eco} , and GPP

A number of environmental factors play an important role in regulating the flux rate of carbon components (NEE, R_{eco} , and GPP) in peatlands (Han et al., 2012). For example, previous studies showed that temperature, light, water table depth, vapor pressure deficient (VPD), biomass, and leaf area index (LAI), mostly controlled the dynamics of CO₂ flux in peatlands (Syed et al. 2006; Fawei et al., 2008; Zhou et al. 2009). ET, VPD, and T_{air} were found the most important drivers on the carbon components in this study (Section 4.3.7). Linear regression analysis is a way to establish whether carbon components could be predicted from easily measured environmental variables. According to single factor regression analysis, Bonneville et al. (2008) reported T_{air} as a good predictor for GPP (linear relationship, $r^2 = 0.55$) and for R_{eco} (exponential relationship $r^2 = 0.54$) but did not account for much of

variability in NEE (linear relationship $r^2 = 0.39$). The similar relationship between T_{air} and R_{eco} was captured by Han et al. (2012). They reported a good linear correlation of NEE, GPP, and R_{eco} with aboveground biomass ($r^2 = 0.87, 0.83$, and 0.7 , respectively). As far as multiple regression analyses were concerned, Zhou et al. (2010) predicted the carbon components of R_{eco} , GPP and NEE as a function of T_{air} , T_{soil} , VPD, and PPFD and reported that predictive powers of GPP model ($r^2 = 0.87$) and R_{eco} ($r^2 = 0.82$) were higher than that of NEE model ($r^2 = 0.52$). When water depth, T_{air} and VPD were factored in the multiple regression models of the carbon components, better predictions were obtained for GPP ($r^2 = 0.65$) and R_{eco} ($r^2 = 0.62$) than for NEE ($r^2 = 0.26$) (Schedlbauer et al., 2012). In this study, multiple regression models were built for each year to predict NEE, GPP, and R_{eco} (Tables 4.14 to 4.17). These model results indicated that variability in GPP, R_{eco} , and NEE were significantly coupled to changes in T_{air} , ET, and VPD. Similar to findings in the related literature, the multiple regression models with a high predictive power were determined for GPP and R_{eco} . The best regression models had r^2 values of 0.61 for NEE, 0.72 for GPP, and 0.78 for R_{eco} which were close to the reported r^2 values above.

5.4. Effects of water fluxes on NEE, R_{eco} , and GPP

The water budget (WB) of the site was investigated according to ET and RF fluxes. RF was always lower than ET (Figure 4.25). The increased difference between RF and ET (i.e. higher water loss from the site) started at spring and went on summer and fall for both in 2011 and 2012 (Figure 4.26). Also, the highest difference occurred in fall. In addition, the difference between RF and ET was higher in 2012 than in 2011 (Figure 4.26). Effects of water budget on carbon components (R_{eco} ,

GPP, NEE) are discussed in the following paragraphs for the measurement period (2011 and 2012).

R_{eco} was directly proportional to ET with a correlation coefficient > 0.5 (Tables 4.11 to 4.14) which indicated that water loss from the peatland increased ecosystem respiration. The water table decline in the peat soil exposes the peat to aerobic conditions, thus leading to increased CO_2 emissions from soils (Juszczak et al., 2013). R_{eco} was highest in 2012 for fall and summer during which the water loss was highest as mentioned above. R_{eco} values during the spring of 2012 were lower than those during the spring of 2011, while water loss was highest in the spring of 2012. This case may be attributed to the flooding duration which lasted longer in 2011 than 2012 (please see Section 5.1).

The correlation matrix showed that ET was directly proportional to GPP ($r > 0.7$) (Tables 4.10 to 4.13) which was consistent with other studies (Brummer et al., 2012 and Jassal et al., 2009). The magnitude of GPP during the growing season (spring and summer) of 2012 was higher than that of 2011 (Table 4.10) parallel to which ET in 2012 was higher than in 2011 (Figure 4.25). A strong positive correlation between ET and biological productivity such as NPP or GPP exists (Chapman et al., 1999).

Lastly, the relationship between NEE and ET had a strong negative correlation ($r > 0.7$). In other words, increased ET led to decreased source strength, which was also reported by Wang et al. (2008). The peatland had a lower growing season cumulative NEE (weaker source strength) (Table 4.9) and a higher growing season ET in 2012 than in 2011 (Figure 4.26).

5.5. Energy balance closure

Reliability of the EC flux data was assessed calculating the closure of surface energy budget (Morrison et al., 2013; Wilson et al., 2002). Theoretically, the ratio of sensible plus latent heat fluxes ($H+LE$) to available energy (R_n-G) should equal to unity (Gao et al., 2009). However, many studies reported energy balance closure ranging from 0.75 to 0.87 (Wilson et al., 2002; Franssen et al., 2010; Barr et al., 2012; Leuning et al., 2012). Many reasons for the incomplete energy closure have been proposed such as sampling errors, bias in instrumentation, neglecting or incorrect accounting for energy storage in soil, and loss of turbulent flux because of low and high frequency (Leuning et al., 2012; Foken et al., 2011; Wilson et al., 2002). The energy balance closure of this study was within 63% for 2011 and 2012 (Figures 4.27 to 4.30). Similarly, Sottocornola et al. (2010) reported 64% for the energy balance closure for four years in a peatland, while other studies reported higher energy balance closure between 80 and 90% in peatlands (Morrison et al., 2013; Lund et al., 2007; Sagerfors, 2007; Corradi et al., 2005). Difficulty with the measurement of ground heat flux (G) was often an important reason for the lack of energy closure (Kurbatova et al., 2002; Corradi et al., 2005) which applied to this study, as well.

The seasonality of the ratio of sensible to latent heat flux (Bowen ratio, β) signals the presence and degree of water stress (limitation) on plant and soil water conditions of the ecosystem which in turn helps to predict seasonal patterns of NEE. β usually has an inverse relationship with SWC (Gu et al., 2006). When plants are free of water stress (high SWC), β is low because most of energy is used as LE by canopy (Wang,

2012). SWC was low when β was high during the non-growing seasons and the converse was true during the growing seasons of 2011 and 2012 in this study (Figure 5.2). The relationship between β and NEE showed a similar pattern during the growing season in 2011 and 2012 (Figure 5.2).

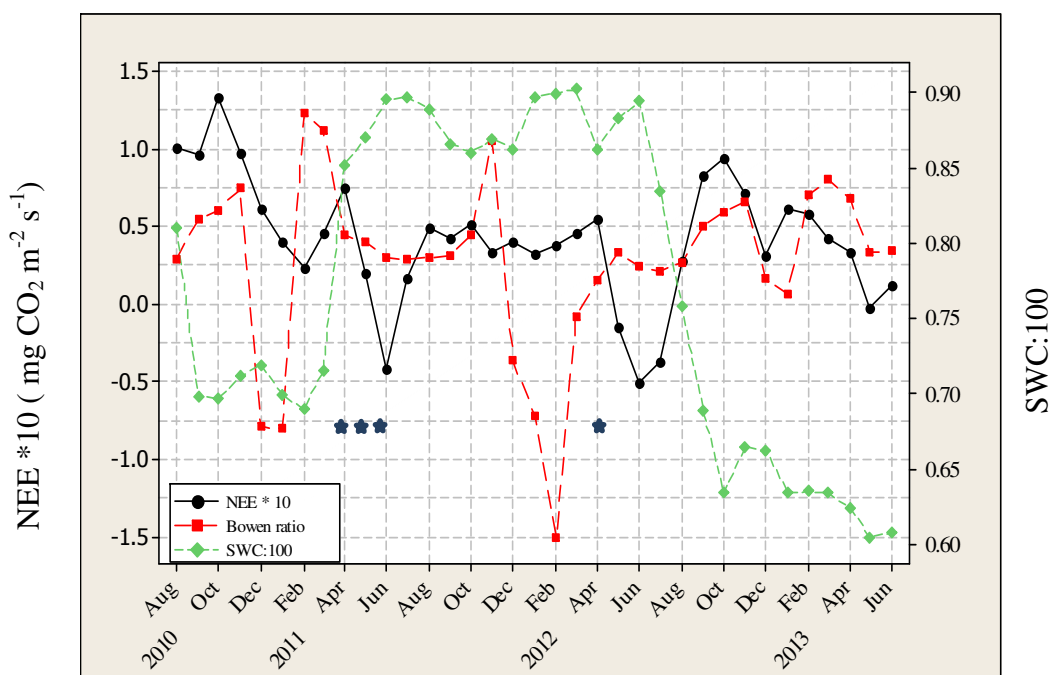


Figure 5.2. Monthly variations in NEE, Bowen ratio, and SWC from August 2010 to June 2013. ★ : Flooding period.

The peatland source strength decreased and became a sink of CO₂ during May to August of 2011 and 2012. β was almost stable between 0.3 and 0.5 during the growing season of 2011 and 2012. The ratio β is reported to typically vary between 0.4 and 0.8 for temperate forests and 2 and 6 for semi-arid areas (Oke, 1992). Strilesky et al. (2012) reported β as 0.5 during July to August for a temperate peatland which was consistent with this study.

5.6. Climate change effects on peatland

Intergovernmental Panel on Climate Change (IPCC) (2007) indicated that CO₂ is by far the most abundant of greenhouse gas (GHG). Simulations performed by the Coupled Model Intercomparison Project Phase 5 (CMIP5) and Earth System Model (ESM) over a spatial resolution of about 1 degree latitude and longitude (about 110 km), which means that finer-scale changes than this resolution cannot be captured, showed that CO₂ concentration would reach 421 ppm under the best-case scenario or 936 ppm under the worst-case scenario by the year 2100 (IPCC 2013). Since the industrial revolution, CO₂ emissions to the atmosphere have contributed significantly to gradual changes in global climate regime including but are not limited to temperature and precipitation. In this dissertation, CO₂ fertilization, increased temperature, and changes in precipitation regime are the main three components of global climate change whose impacts on peatland ecosystems are focused on.

Increased atmospheric CO₂ concentration has the potential to boost vegetation productivity in the absence of other limiting factors, an effect known as CO₂ fertilization ((Parish et al., 2008; Matthews, 2007). CO₂ fertilization affects plants differently depending on photosynthetic pathways (C3 versus C4) (Ehleringer et al., 2005). When other environmental resources such as light, water, and nutrients are not limited, C3 species grown under elevated CO₂ conditions have shown enhanced photosynthesis (Tang et al., 2009). Conversely, C4 species have been found to be generally less responsive to elevated CO₂ levels than C3 species (Lee, 2011). However, the rate of photosynthesis would increase with higher CO₂ concentrations only up to a saturation point, beyond which photosynthesis remains constant

(Houghton et al. 2001). Saturation occurs at ambient CO₂ concentrations of about 400 ppm for C₄ plants and above 600 ppm for C₃ plants (Bonan, 2008).

The globally averaged surface temperature is anticipated to increase by 0.3 to 4.8°C by the period of 2081 to 2100, when compared to the 1986 to 2005 period (IPCC, 2013). The climate scenarios have a lower confidence level for projected changes in precipitation than in temperature (IPCC 2007). IPCC (2013) reported that mean annual precipitation is likely to increase in the high latitudes, mean precipitation will likely to decrease in many mid-latitudes and subtropical dry regions, while mean precipitation will decrease in many mid-latitude wet regions. In addition, extreme precipitation events will be very likely to become more intense in most of the mid-latitudes and wet tropical regions (IPCC 2013). Similar to IPCC (2013), Strack (2008) reported that global average precipitation will exhibit regional variability. It is very likely that heavy precipitation events will become more frequent in high-latitudes (Belyea et al., 2004). Projections for the mid-latitudes projections suggest an increase in precipitation in winter and a decrease in some peatland regions in summer (e.g. Western Europe) (Parish et al., 2008). Summer precipitation is also projected to decrease in the Mediterranean basin with an increased risk of drought (Parish et al. 2008).

Peatlands as a globally important stock of soil organic carbon (Gorham 1991; Dise 2009) are most likely to be one of the most sensitive ecosystems to climate change (IPCC, 2007). The assessment of future effects of elevated CO₂ concentrations on peatland carbon sequestration is rather difficult, because any response to increased atmospheric CO₂ will be offset by other concurrent biogeochemical and hydrological

changes. Direct and indirect effects of increased temperatures under climate change such as increased decomposition and mineralization of soil organic carbon, and lowered groundwater table would be crucial to sustainability of peatlands (Sulman, 2009). Given the domino effects on ecosystem structure and function and the complexity of interactions among a whole host of factors, air temperature increase is difficult to assess in terms of whether or not to augment the sink or source strength of the peatlands (Parish et al., 2008). The dominance of NPP versus R_{eco} under increased temperature would control the fate of the peatlands to serve as a sink or source of CO_2 (Zhaojun et al., 2011). Increased temperature together with decreased precipitation may expose the peatlands to anaerobic conditions which in turn increases decomposition and mineralization of soil organic matter (CO_2 emissions by soil respiration to the atmosphere) (Lloyd, 2006).

Changes projected in relation to climate change for Turkey include an increase in mean annual temperature and a decrease in precipitation especially in the west and south regions of the country (Dalfes et al., 2007; Önoğlu and Semazzi, 2009). According to Turkey's National Climate Change (2010) report, surface temperature is projected to increase by about $3.5^{\circ}C$ in winter and $6^{\circ}C$ in summer by the end of the 21st century. The same report stated that precipitation is expected to decrease by up to 70% in summer, 23% in fall, 35% in winter, and 47% in spring by the end of 21st century. Under these conditions, increased CO_2 emissions to the atmosphere are anticipated as a result of increases in temperature and the frequency of summer dry periods with decreases in groundwater table depth and soil moisture. Some studies showed that peatlands are projected to change from a sink to a source of CO_2 with increased temperature and decreased soil water content (e.g. Tarnocai 2006; Li et al.,

2007; Dorrepaal et al., 2009; Fan et al., 2012). However, response of peatlands to climate change still remains to be explored to better understand their role in global carbon cycle and to better devise mitigative and preventive measures towards their sustainability (Limpens et al., 2008; Zhaojun et al., 2011).

CHAPTER VI

6. CONCLUSIONS

CO₂, energy, and water exchanges of a cool temperate Yeniçağa peatland were estimated using eddy covariance (EC) flux data from August of 2010 to July of 2013. Answers to the initial research hypotheses and questions in the Introduction section are summarized below.

- Is Yeniçağa peatland a CO₂ sink or source over the measured time period?

The peatland acted as strong C source on an annual basis. Extreme events observed (flooding and drought periods) during the measurement period caused an increase in the source strength of CO₂. Above all, the main driver for the peatland to behave as a strong source of CO₂ is most likely to be the human-induced disturbances such as peat mining, drainage, clearance for agricultural and recreational land uses, and overgrazing.

- What are GPP and R_{eco} flux components of NEE? In other words, how much CO₂ is used by GPP and released by R_{eco} by Yeniçağa peatland?

The annual assimilation of CO₂ by gross primary productivity (GPP) was estimated at 944.9 g C m⁻² in 2011 and 1010.7 g C m⁻² in 2012. Release of CO₂ emissions to the atmosphere by ecosystem respiration (R_{eco}) amounted to 1228.6

g C m⁻² in 2011 and 1256.1 g C m⁻² in 2012. The cumulative net ecosystem exchange (NEE) was found as 283.7 and 245.4 g C m⁻² for 2011 and 2012 because of higher R_{eco} rates. Relationships among carbon components indicated that there was a stronger correlation between NEE and GPP ($0.81 < r < 0.85$) than between NEE and R_{eco} ($0.14 < r < 0.27$) which showed that variability in GPP primarily controls NEE.

- What are the main environmental controls over NEE, GPP and R_{eco} ? Is it possible to model temporal dynamics of NEE and its components as a function of environmental variables?

Multiple linear regression analysis clearly revealed that variability in GPP, R_{eco} , and NEE was significantly coupled to changes in T_{air} , ET, and VPD with a high predictive power for GPP and R_{eco} . The best regression models had r^2 values of 0.61 for NEE, 0.72 for GPP, and 0.78 for R_{eco} .

- What are temporal dynamics of water fluxes over the measurement period?

Water budget (WB) of Yeniçağa peatland showed a state of water deficit for the ecosystem as evapotranspiration (ET) was higher than rainfall (RF). For this reason, the peatland was exposed to anaerobic conditions which in turn increased CO₂ release by soil respiration to the atmosphere.

- Do energy fluxes comply with energy balance closure?

The ratio of H+LE to R_n-G should be unity according to the energy balance closure, and it was found to be 0.63 for 2011 and 2012 in this study. The lack of energy balance closure may be attributed to the incorrect accounting of soil heat flux (G). Bowen ratio (β) (the ratio of H to LE) as an aridity indicator of an

ecosystem revealed the peatland can be characterized as a mesic temperate ecosystem during the measurement period.

Increased temperature and decreased precipitation as projected for the climate change scenarios of Turkey may have significant consequences for the peatland ecosystems, especially, for an already strong source character of Yeniçağa peatland. Decreased precipitation and prolonged drought period appear to enhance the source strength of the peatland unless preventive and mitigative measures are taken against the disturbances of Yeniçağa peatland and towards its sustainable management.

Flux tower-based long-term measurements of CO₂ and H₂O fluxes can be related to remotely-sensed observations including satellite and radar images, and aerial photos by (un)manned air vehicles. Interrelating eddy covariance data and remotely-sensed data poses opportunities for up-scaling of CO₂ and H₂O dynamics not only to broader spatial scales such as watersheds at which land-use planning- and management-related decisions are most robust to be made but also to complex and/or remote terrains such as mountains where flux towers are impossible to construct. Integration of geographical information systems, remote sensing techniques, image processing, denoising algorithms, empirical models, process-based models, and inclusion of socio-economic variables remains to be explored in the future research directions. Thus, we can come up with ways in which changes in inventories of C sources (emissions) and sinks (removals) are monitored everywhere and every time across Turkey or the globe in response to human-induced disturbances and act accordingly and timely in terms of taking preventive and mitigative measures in the face of changing global environment and climate.

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