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## Dynamics of water vapor and energy exchanges above two contrasting Sudanian climate ecosystems in Northern Benin (West Africa)

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## RESEARCH ARTICLE

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## Key Points:

- At least 70% of the available energy was changed into actual evapotranspiration during wet seasons
- Soil moisture was found to be the main factor controlling energy partitioning
- The annual cumulative AET measured was 30% higher over the woodland than over the cultivated area

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## Dynamics of water vapor and energy exchanges above two contrasting Sudanian climate ecosystems in Northern Benin (West Africa)

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**Abstract** Natural ecosystems in sub-Saharan Africa are experiencing intense changes that will probably modify land surface feedbacks and consequently the regional climate. In this study, we have analyzed water vapor ( $Q_{LE}$ ) and sensible heat ( $Q_H$ ) fluxes over a woodland (Bellefoungou, BE) and a cultivated area (Nalohou, NA) in the Sudanian climate of Northern Benin, using 2 years (from July 2008 to June 2010) of eddy covariance measurements. The evaporative fraction (EF) response to environmental and surface variables was investigated at seasonal scale. Soil moisture was found to be the main environmental factor controlling energy partitioning. During the wet seasons, EF was rather stable with an average of  $0.75 \pm 0.07$  over the woodland and  $0.70 \pm 0.025$  over the cultivated area. This means that 70–75% of the available energy was changed into actual evapotranspiration during the investigated wet seasons depending on the vegetation type. The cumulative annual actual evapotranspiration (AET) varied between  $730 \pm 50 \text{ mm yr}^{-1}$  at the NA site and  $1040 \pm 70 \text{ mm yr}^{-1}$  at the BE site. With similar weather conditions at the two sites, the BE site showed 30% higher AET values than the NA site. The sensible heat flux  $Q_H$  at the cultivated site was always higher than that of the woodland site, but observed differences were much less than those of  $Q_{LE}$ . In a land surface conversion context, these differences are expected to impact both atmospheric dynamics and the hydrological cycle.

### 1. Introduction

Natural ecosystems in sub-Saharan Africa have undergone intense changes over the last few decades. Between 1975 and 2000, the area of agricultural lands increased by 57% [Eva *et al.*, 2006], at the expense of forests which formerly covered 33% of the region [Food and Agriculture Organization, 2011; Ouedraogo *et al.*, 2010]. This change was driven by high population growth [DeFries *et al.*, 2010; Guengant and Kamara, 2012], about 2.7% per year during the 1975–2000 period [Eva *et al.*, 2006; Heldmann and Doevenspeck, 2009], and expected to reach 4% per year by 2050 according to the most recent United Nation's projection [UN, 2015].

It is well known that land cover changes affect the net radiation partitioning into sensible heat and water vapor fluxes and consequently affect global atmospheric circulation [Miralles *et al.*, 2011; Numaguti, 1993; Pielke *et al.*, 2002] and the planetary boundary layer through land-atmosphere feedbacks [Xue *et al.*, 2004]. These feedbacks have been identified to be particularly strong in West Africa where the surface moisture distribution drives both monsoon and squall line dynamics [Charney, 1975; Koster *et al.*, 2004; Taylor *et al.*, 2012]. Vegetation changes that impact the continental water cycle [Amogu *et al.*, 2015; Descroix *et al.*, 2012, 2009] are also expected to significantly affect moisture convergence, precipitation [Wang and Eltahir, 2000], and consequently the regional climate along with the overall hydrological cycle [Alo and Wang, 2010; Weaver and Avissar, 2001]. Therefore, studies aimed at analyzing and quantifying the water vapor and sensible heat fluxes over several ecosystems in West Africa are necessary to investigate how the observed land cover changes will impact these feedbacks under changing climatic conditions.

Several studies have focused on land surface fluxes in temperate [Aubinet *et al.*, 2000; Wilson *et al.*, 2002] as well as boreal climates [Baldocchi *et al.*, 2004; Baldocchi and Ryu, 2011]. However, much less attention has been paid to the West Africa regions because of the scarcity of observations data. In this region, water vapor and energy fluxes of the Sahelian region were first analyzed after the pioneering Hapex-Sahel experiment covering a 2 months observation period [Gash *et al.*, 1997]. More recently, longer periods have been analyzed

for various vegetation covers [Brümmer *et al.*, 2008; Ramier *et al.*, 2009; Timouk *et al.*, 2009] leading to the first complete validation of land surface model simulations under Sahelian climate conditions [Velluet *et al.*, 2014].

Few studies are however available for the Sudanian climate. A first study was conducted with a scintillometer over heterogeneous vegetation [Guyot *et al.*, 2009]. The studied area was about 1 km<sup>2</sup> and included a mix of cultivated land, woodland, and a riparian forest. Although this study provided useful insights at landscape scale, no conclusions could be drawn on the contribution of the different land covers. Recently, Mamadou *et al.* [2014] analyzed water vapor and energy fluxes over 1 year (2008) above a cultivated area. This area was covered by crops alternating with herbaceous fallows [Mamadou *et al.*, 2014]. The authors pointed out (i) contrasted surface responses depending on the season and (ii) the importance of surface temperature, which directly drives the longwave radiation budget and sensible heat fluxes and indirectly drives water vapor fluxes through transpiration limitations. Until now, no water vapor and sensible heat flux observations have been published for woodlands in the Sudanian region, despite their potential role in modifying land surface feedbacks to the atmosphere and their role in favoring infiltration and soil conservation.

This study is the first to compare surface exchanges over woodland and cultivated area in the Sudanian climate based over a long observation period (2 years). It analyzes the seasonal dynamics of water vapor and energy exchanges and their diurnal cycles by means of the eddy covariance (EC) technique. Both study sites are part of the AMMA-CATCH long-term observation system ([www.amma-catch.org](http://www.amma-catch.org)) aimed at studying the impact of climate and land surface changes on the hydrological cycle [Lebel *et al.*, 2009]. In this study, we will (i) quantify the differences and identify similarities in the water vapor and energy exchanges for different land covers, (ii) identify the key factors controlling energy partitioning at the seasonal scale, (iii) evaluate the annual actual evapotranspiration (AET) and the associated uncertainties, and finally (iv) discuss the implications of land cover changes for both land surface feedbacks and the water budget under a deforestation context.

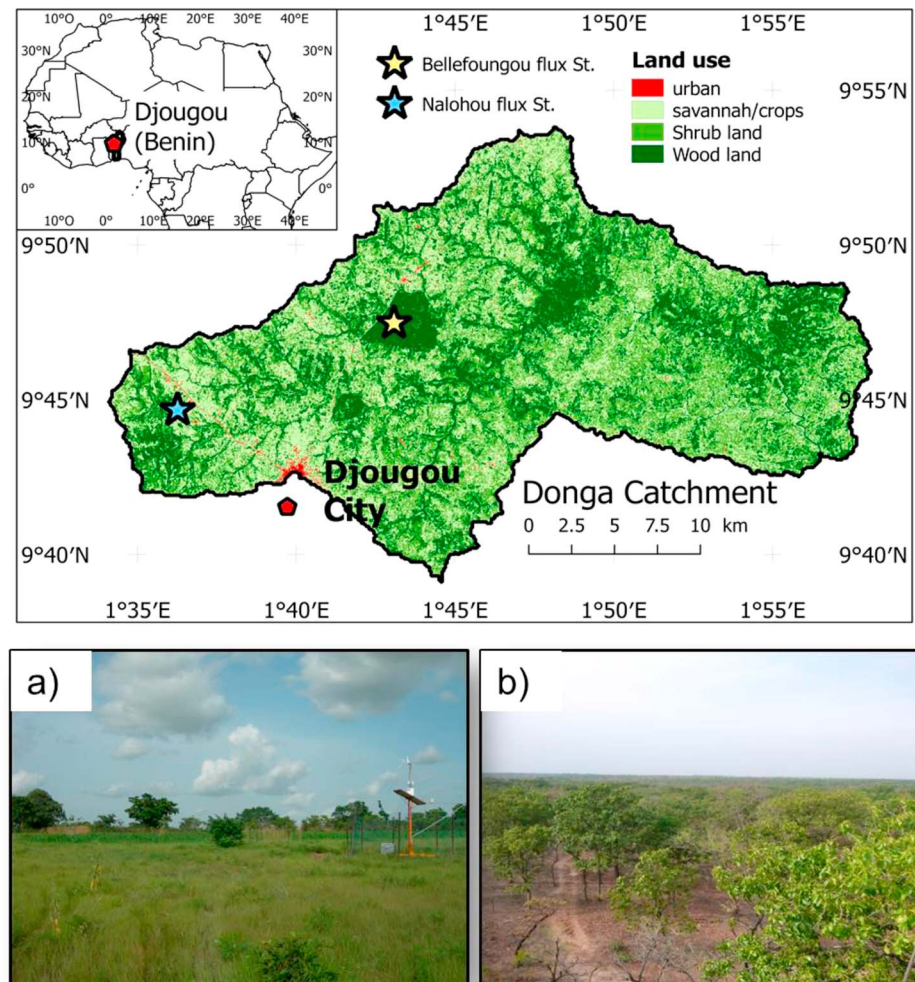
## 2. Materials and Methods

### 2.1. Study Sites

The measurements were conducted over two contrasting sites located in the Donga catchment that extends over ~600 km<sup>2</sup> in northern Benin, 450 km northwest of Cotonou (Figure 1). Vegetation is a patchwork of woodland (47%), shrubland (31%), and a mixture of crops and herbaceous fallow (22%) composed mainly of perennial grasses [Seghieri *et al.*, 2009] (Figure 1). According to the World Reference Base (IUSS Working Group WRB, 2006), the soils are mainly ferric lixisols [Faure and Volkoff, 1998; Giertz and Diekkrüger, 2003]. The first soil horizon A (40 cm in the cultivated areas and 60 cm in the woodland) is a loamy sand with a sandy loam underneath (U.S. Department of Agriculture (USDA) soil classification). The underlying red-orcher B horizon is a sandy clay loam, the C horizon lies deeper than 1.6 m. This profile is consistent with numerous tropical soils [de Condappa *et al.*, 2008]. The landscape is relatively flat overall, with gentle slope and elevation ranging between 450 and 550 m.

The two study sites are located 13 km apart in the Donga catchment. The cultivated area of Nalohou (hereafter Nalohou (NA)) is located near Nalohou village (latitude 9.74°N, longitude 1.60°E, 449 m asl). The landscape is relatively flat with local slopes not exceeding 3%. The Nalohou site is a typical cultivated area of the region, including crops and herbaceous fallow, and is dominated by C4 plants. The NA flux tower is located in a small herbaceous fallow parcel, surrounded by rotating crops (yam, maize, cassava, and groundnuts). The flux footprint is calculated with the analytical Hsieh 1-D model [Hsieh *et al.*, 2000] with a 2-D extension [Detto *et al.*, 2006]. It covers about 4000 m<sup>2</sup> during the day and more during the night, and was representative of cultivated areas of the region, including fallows (see Mamadou *et al.* [2014] for the complete footprint analysis).

The second site is located in a woodland, the most widespread vegetation type in tropical Africa [White, 1983], near Bellefoungou village (latitude 9.791°N, longitude 1.718°E, 445 m asl) (hereafter Bellefoungou (BE)). Vegetation of this woodland consists in an open stand of trees with crown heights of around 15 m (Figure 1b), covering 60%–80% of the total soil area [Ago *et al.*, 2016]. The canopy casts little shade and an annual herbaceous layer, composed of perennial C4 vegetation which varies from 0 (dry season) to 50 cm (wet season) covers the ground. The dominant overstory species are mainly composed of C3 vegetation including: (1) *Isobерlinia tomentuosa*, (2) *Isobерlinia doka* and *Burkea Africana*, and (3) *Vitellaria paradoxa* [Houéto *et al.*, 2012; Seghieri *et al.*, 2009]. This floristic composition defines a *isobерlinia* woodland. The



**Figure 1.** Locations of (a) Nalohou (cultivated area) and (b) Bellefougou (woodland) flux stations in the Donga Catchment in Northern Benin. Shaded areas display the main land cover categories.

canopies of *Isoberlinia doka*, the dominant species at the BE site, shed at least 50% of its leaves but never 100% between November and February [Seghieri *et al.*, 2012]. This species composition is reasonably homogeneous around the BE flux tower and contributes to the observed footprint (about 60,000 m<sup>2</sup>) whatever the season [Mamadou, 2014]. The mean local slope is 2.5%.

The region is subjected to a Sudanian climate, driven by the fluctuation of the Intertropical Convergence Zone [Sultan and Janicot, 2003], is characterized by a succession of wet and dry seasons separated by two transitional periods. Mamadou *et al.* [2014] have shown that the absolute humidity ( $q_a$ ) can be used to delimit the seasons. The dry ( $q_a < 6 \text{ g m}^{-3}$ ) and wet ( $q_a > 16 \text{ g m}^{-3}$ ) seasons are separated by two transitional periods during which the absolute humidity gradually varies. The mean precipitation amount in the region is 1230 mm over the period 1950–2010. The distribution of precipitation is unimodal and more than 70% of the annual rainfall is received after the monsoon onset, between July and October [Lelay and Galle, 2005].

## 2.2. Study Period and Its Characteristics

Results presented in this paper cover two hydrological years extending from July 2008 to June 2010. The annual precipitation of the investigated years was above the mean annual rainfall: 1364 mm (2008), 1467 mm (2009), and 1534 mm (2010), respectively. According to the season definitions proposed by Mamadou *et al.* [2014], the studied periods include two “half” wet seasons (2008 and 2010), one complete wet season (2009) and two subsequent dry seasons (2008–2009 and 2009–2010). The dry seasons were short:

**Table 1.** Instrumentation of Nalohou (Cultivated Area) and Bellefoungou (Woodland)

Instrument	Sampling Rate	Nalohou Sensors and Manufacturers	Height (m)	Bellefoungou Sensors and Manufacturers	Height (m)
3-D sonic anemometer	20 Hz	CSAT3 Campbell Sci., USA	5	CSAT3 Campbell Sci., USA	18
Gas analyzer	20 Hz	LI-7500, Li-COR Bioscience, USA	5	LI-7500, Li-COR Bioscience, USA	18
Thermometer Hygrometerr	10 s.	WXT510, Vaisala, Finland	2	WXT510, Vaisala, Finland	5
Pyranometer	10 s.	CNR1, Kipp and Zonen, NL	2	SKS1110, Skye, NL	5
Net radiometer	10 s.	CNR1, Kipp and Zonen, NL	2	NR-Lite, Kipp and Zonen, NL	5
Raingauge	pulse	Tipping bucket ABS3030, Précis Mécanique, France	1.2	Tipping bucket ABS3030, Précis Mécanique, France	1.2
Soil thermometer	10 s.	Thermocouple 105, Campbell Sci., USA	−0.5, −0.1, −0.2, −0.4, −0.6, −1.0	Thermocouple 105, Campbell Sci., USA	−0.5, −0.1, −0.2, −0.4, −0.6, −1.0
Water content probe	10 s.	CS616, Campbell Sci., USA	−0.5, −0.1, −0.2, −0.4, −0.6, −1.0	CS616, Campbell Sci., USA	−0.5, −0.1, z −0.2, −0.4, −0.6, −1.0
Data logger		CR3000, Campbell Sci, USA		CR3000, Campbell Sci, USA	

31 and 41 days, characterized by northern Harmattan winds. The wet seasons lasted about 120 days, dominated by monsoon moisture blowing from the southwest and accompanied by heavy rainy events. Between these two seasons, drying (resp. moistening) transitional periods lasted 68 and 56 days (resp. 78 and 73 days).

### 2.3. Measurement Systems

Turbulent fluxes were measured continuously, from July 2008 to June 2010, using the eddy covariance (EC) system. The EC sensors (Table 1) were installed on a mast at a height of 5 m at NA and 18 m at BE, respectively.

At both sites, a complete meteorological station measuring rainfall, wind speed and direction, air temperature, relative humidity, air pressure, and a complete radiative budget was also installed (Table 1). Soil measurements included the soil moisture and the soil temperature profiles down to 1 m deep (Table 1). Meteorological data and soil measurements were recorded at a 30 min time step.

The leaf area index (LAI) at daily time step was derived from a SEVIRI satellite time series ( $3 \times 3 \text{ km}^2$ ), constrained by in situ data derived from hemispherical photographs using the method proposed by Weiss *et al.* [2004].

### 2.4. Energy and Water Vapor Flux Computations and Data Processing

Half-hourly turbulent fluxes were computed from the covariance (equations (1) and (2)) of the vertical wind component fluctuation  $w'$  ( $\text{m s}^{-1}$ ) with those of the scalar quantities, namely, air temperature  $T'$  (K) for the sensible heat flux, and absolute humidity  $q'$  ( $\text{g m}^{-3}$ ) for the water vapor flux.

$$Q_H = \rho C_p \overline{w'T'}, \quad (1)$$

$$Q_{LE} = \lambda \overline{w'q'}, \quad (2)$$

where  $\rho$  is the air density ( $\text{kg m}^{-3}$ ),  $C_p$  the heat capacity at constant pressure ( $\text{J kg}^{-1} \text{K}^{-1}$ ), and  $\lambda$  the latent heat of vaporization ( $\text{J g}^{-1}$ ). Fluxes were computed with EdiRe software version 1.5.0.28 (University of Edinburgh, UK) using the standard procedures defined in the CarboEurope protocol [Aubinet *et al.*, 2000]. Further details on data processing can be found in Mamadou *et al.* [2014]. Half-hourly  $Q_H$  and  $Q_{LE}$  data for which the quality tests [Mauder and Foken, 2004] were not satisfied were rejected from the data set. Eddy fluxes corresponding to the rainy events and the half-hour which followed the rainfall were also removed from the data set to avoid artifacts due to the high sensitivity of the hygrometer to raindrop extinction [Culf *et al.*, 2004].

After postprocessing and filtering, 76% and 62% of  $Q_H$  and 63% and 48% of  $Q_{LE}$  at NA, and, 71% and 64% of  $Q_H$  and 64% and 40% of  $Q_{LE}$  at BE, respectively, during the 2008–2009 and 2009–2010 hydrological years, were considered suitable for further analysis.

Before filling the gaps in the data, the annual energy balance closure was evaluated for each site using the linear regression of the sum of half-hourly turbulent fluxes against the available energy (equation (3)).

$$Q_H + Q_{LE} = Q - Q_G - Q_{\Delta S} \quad (3)$$

with

$$Q_{\Delta S} = S_H + S_{LE} = \int_0^z \rho C_p \frac{dT}{dt} dz + \int_0^z \lambda \frac{dq}{dt} dz, \quad (4)$$

where  $Q$  is the net radiation and  $Q_G$  the soil heat flux estimated by the harmonic method from soil temperature and moisture profiles [Guyot *et al.*, 2009].  $Q_{\Delta S}$  is the energy storage in the air column beneath the eddy covariance system level  $z$  (m) [Barr *et al.*, 2006] and was found (not presented here) to be negligible for both sites, even at the 15 m high woodland site (less than  $15 \text{ W m}^{-2}$ ). All terms in equation (3) have units of ( $\text{W m}^{-2}$ ).

Ideal energy balance closure is represented by a slope of 1 and origin of zero for a plot of equation (3). For the presented data set, the average slope over the studied years was 0.84 ( $r^2 = 0.92$ ) at NA and 0.88 ( $r^2 = 0.82$ ) at BE. Annual values of the energy balance ratios were thus calculated as  $\sum (Q_H + Q_{LE}) / \sum (Q - Q_G - Q_{\Delta S})$  and were, respectively, 0.77 at NA and 0.90 at BE. These nonnegligible biases fall within the range reported for various ecosystems [Wilson *et al.*, 2002] and are classic for the EC technique [Foken *et al.*, 2011].

## 2.5. Statistical Analysis and Derived Quantities

### 2.5.1. Calculation of Midday Averages and Diurnal Cycles

Midday averages (1000–1400 UTC) of the “measured turbulent fluxes” were calculated to analyze the seasonal cycle. Indeed, more than 80% of  $Q_H$  and  $Q_{LE}$  data were available within that time slot and the dynamics were representative of the daily average values [Mamadou, 2014]. The evaporative fraction (EF) was also calculated using midday average fluxes.

$$EF = \frac{Q_{LE}}{Q_{LE} + Q_H}. \quad (5)$$

The diurnal cycle of energy balance components over the two sites was also characterized during the different periods of the monsoon cycle (dry and wet seasons and transitional moistening and drying periods) for the year 2009. With this objective, half-hourly data of energy balance components were averaged to build a specific composite daily cycle for each season.

### 2.5.2. Calculation of Bulk Parameters

The surface conductance  $G_s$  ( $\text{m s}^{-1}$ ) quantifies the ability of the soil or the canopy to transfer water vapor to the atmosphere. The midday average of  $G_s$  was calculated by inversion of the Penman-Monteith equation as follows:

$$G_s = \left[ \frac{1}{G_a} \left( \frac{\Delta}{\gamma} \beta - 1 \right) + \frac{\rho C_p \text{VPD}}{\gamma (R_n - G)} (\beta + 1) \right]^{-1}, \quad (6)$$

where  $\Delta$  ( $\text{Pa K}^{-1}$ ) is the slope of the saturation curve,  $\gamma$  ( $\text{Pa K}^{-1}$ ) the psychrometric constant, VPD (Pa) the vapor pressure deficit,  $\beta$  the Bowen ratio, defined as  $Q_H/Q_{LE}$ , and  $G_a$  ( $\text{m s}^{-1}$ ) the aerodynamic conductance, is the ability of the atmospheric surface layer to transfer water vapor from the air near the vegetation to the atmosphere, was calculated using the following equation as proposed by Mamadou *et al.* [2014]:

$$G_a = \frac{u_*}{u} \frac{\phi_H}{\phi_m}, \quad (7)$$

where  $u^*$  ( $\text{m s}^{-1}$ ) is the friction velocity,  $u$  ( $\text{m s}^{-1}$ ) the wind speed, and  $\frac{\phi_H}{\phi_m}$  are the Businger-Dyer stability function [Businger *et al.*, 1971].

Finally, the midday average of decoupling coefficient ( $\Omega$ ) [Jarvis and McNaughton, 1986] was computed as

$$\Omega = \frac{\Delta/\gamma + 1}{\Delta/\gamma + 1 + G_a/G_s}. \quad (8)$$

( $\Omega$ ) varies between 0 and 1. Values approaching zero indicate that  $Q_{LE}$  is highly sensitive to surface conductance and relative water vapor deficit, while those approaching 1 indicate that  $Q_{LE}$  is mostly controlled by net radiation [Jarvis and McNaughton, 1986].

For statistical analysis, the soil moisture profile was integrated to estimate the soil water content in the superficial layer (0–30 cm) and in the first meter of the soil (0–100 cm) and then normalized to be compared from one site to another. The extractable water  $\chi$  was calculated as follows:

$$\chi = \left( \frac{\theta - \theta_{\min}}{\theta_{\max} - \theta_{\min}} \right), \quad (9)$$

where  $\theta$  ( $\text{m}^3 \text{m}^{-3}$ ) is the daily average of integrated soil moisture and  $\theta_{\min}$ ,  $\theta_{\max}$  ( $\text{m}^3 \text{m}^{-3}$ ) the annual minimum and maximum of soil moisture. Finally, Spearman partial correlation between midday evaporative

fraction and its environmental and surface variables was carried out to identify the main controls of energy partitioning at seasonal scale.

### 2.6. Annual AET and Uncertainty Estimations

In order to estimate the annual actual evapotranspiration (AET), small gaps (i.e., < 2 h) in turbulent flux data ( $Q_{LE}$ ) were filled by linear interpolation over time. Then the remaining data gaps in half-hour  $Q_{LE}$  data were filled using two approaches. Nighttime (incoming shortwave radiation lower than  $10 \text{ W m}^{-2}$ ) gaps were replaced by the nighttime average seasonal values of  $Q_{LE}$ , whereas daytime gaps, if any, were filled using the linear relationships between retained values of  $Q_{LE}$  as the dependent variable and measured net radiation ( $Q$ ) as the independent variable. This method assumes that EF may be considered as constant during each season. This seasonal  $Q_{LE}/Q$  fraction was relatively constant during dry and wet seasons. This was not the case during intermediate periods, but few daily values were missing during these periods. The  $Q_{LE}$  fraction of  $Q$  was determined during the different monsoon periods using linear regression with the origin forced through zero. Finally, longer gaps up to 5 days were filled using the mean diurnal variation of the variable [Falge *et al.*, 2001].

The measured water vapor fluxes are affected by uncertainties due to the presence of both random ( $\delta_{ran}$ ) and systematic ( $\delta_{sys}$ ) errors. Systematic errors are difficult to estimate [Billesbach, 2011; Hollinger and Richardson, 2005]. The energy balance closure obtained ( $>0.80$ ) in this study indicates that the systematic errors in the turbulent fluxes could be 20% if both  $Q_H$  and  $Q_{LE}$  were impacted similarly. This is intrinsic to the EC method and not considered in this study as we aim to compare two data series determined using this methodology. The remaining systematic errors were attributed to the biases due to gap filling ( $\delta_{gap}$ ), estimated for both daytime and nighttime fluxes. Indeed, for the nighttime fluxes, they were estimated as the standard errors of the average nighttime seasonal values of non-gap-filled  $Q_{LE}$ , whereas for the daytime fluxes, they were estimated as the standard errors of the linear regression residuals. The random error ( $\delta_{ran}$ ) in each half-hour flux was also estimated using the relations established by [Richardson *et al.*, 2006] for grassland and forest ecosystems. The total uncertainty in the gap-filled AET,  $\delta(\text{AET})$ , was finally computed at daily and annual scales by summing these three error sources as indicated in equation (10) [Hollinger and Richardson, 2005]. All statistical calculations (Spearman partial correlation coefficients, daily averages, and associated errors), models, and figures were made using R statistical software (R version 3.1.1).

$$\delta(\text{AET}) = \sqrt{(\Sigma\delta_{ran}^2 + \Sigma\delta_{gap\_nighttime}^2 + \Sigma\delta_{gap\_daytime}^2)}. \quad (10)$$

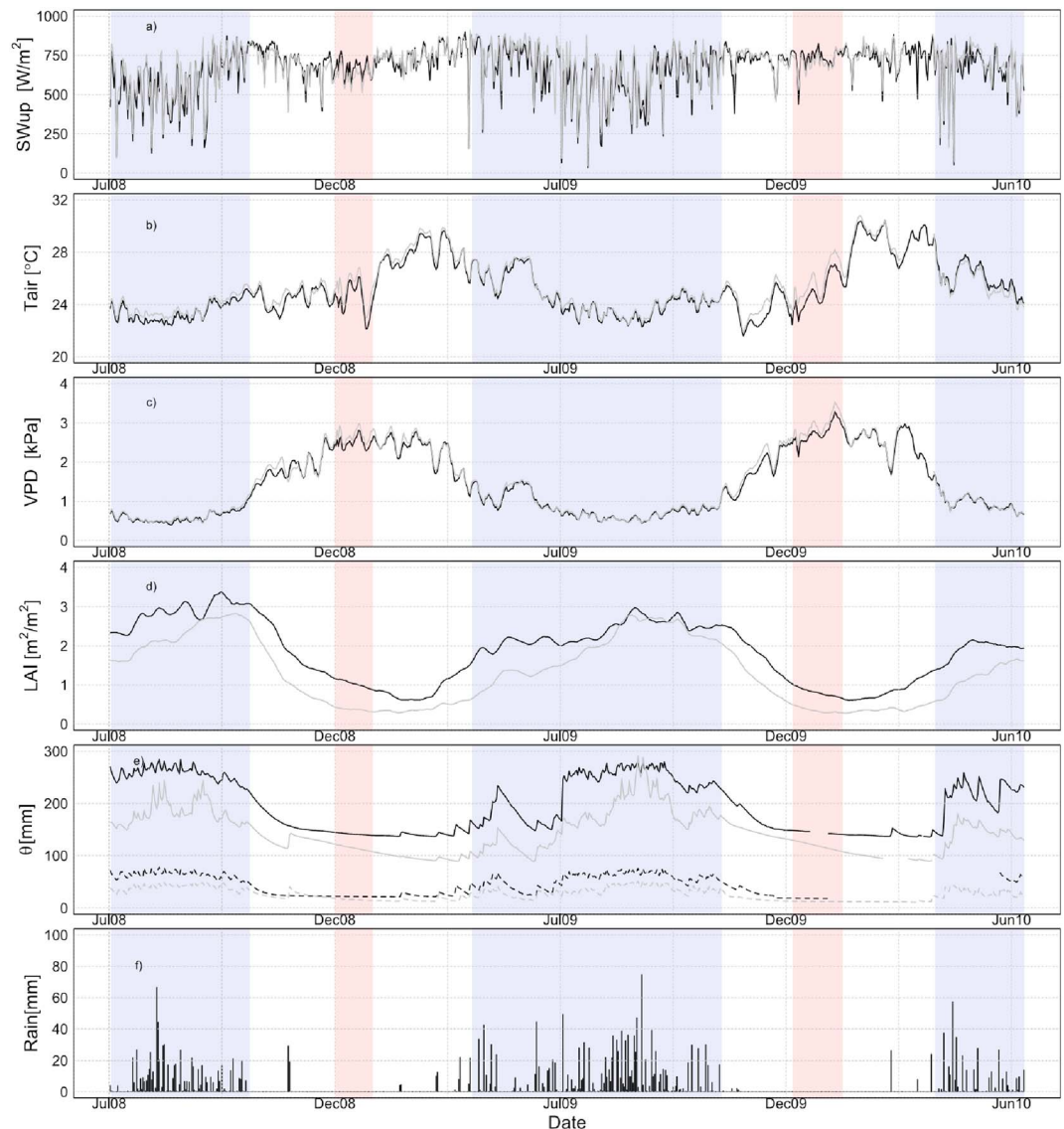
## 3. Results

### 3.1. Meteorological and Surface Conditions

The annual cycle of incoming shortwave radiation ( $SW_{up}$ ), air temperature ( $T_{air}$ ), water vapor pressure deficit (VPD), soil water content ( $\theta$ ) at the 0–30 cm and 0–1 m layers, and leaf area index (LAI) at NA and BE are shown in Figure 2. Blue and pink shading indicate wet and dry seasons defined with the absolute humidity criteria [Mamadou *et al.*, 2014].

The daily meteorological conditions of the two sites, located 13 km apart, were expectedly close except for LAI and soil moisture. At both sites, minimum annual values of  $SW_{up}$  were observed during the wet seasons mainly due to cloud cover [Gounou *et al.*, 2012; Kounouhéwa *et al.*, 2013]. During the dry seasons,  $SW_{up}$  was lower than its value during the moistening period (Figure 2a) due to the aerosols which are brought by the Harmattan winds in the region [Marticorena *et al.*, 2010]. The VPD showed a seasonal cycle with a yearly maximum of 3 kPa and minimum of 0.8 kPa during dry and wet seasons, respectively (Figure 2c); suggesting a high atmospheric demand for water vapor fluxes. The yearly maximums of  $T_{air}$  ( $\sim 30^\circ\text{C}$ ) were obtained during the moistening period prior to wet seasons while the minimums were observed in the wet season (Figure 2b). These daily values hide hot temperature conditions which are able to limit photosynthesis and therefore evapotranspiration as showed by Mamadou *et al.* [2014].

The seasonal variations of  $\theta_{0-30 \text{ cm}}$  and  $\theta_{0-1 \text{ m}}$  were consistently greater at BE than at NA (Figure 2e), but the magnitudes were similar. The increase of soil water content was earlier and faster in the woodland than on the cultivated area. This is probably due both to a lower evaporation under the shade of trees at the beginning of the season and to the roughly 10% higher rainfall observed on the woodland site during the studied years.



**Figure 2.** The 5 day average of: (a) midday incoming shortwave radiation ( $SW_{up}$ ), (b) daily air temperature ( $T_{air}$ ), (c) daily vapor pressure deficit (VPD), (d) daily leaf area index (LAI), (e) daily soil water content ( $\theta$ ) in the 0–30 cm (dotted line) and 0–1 m (solid line) layers, and (f) the daily sums of precipitation at Nalohou (gray line) and Bellefoungou (black line) during the studied years. The pink and blue shaded areas indicate, respectively, the dry and wet seasons.

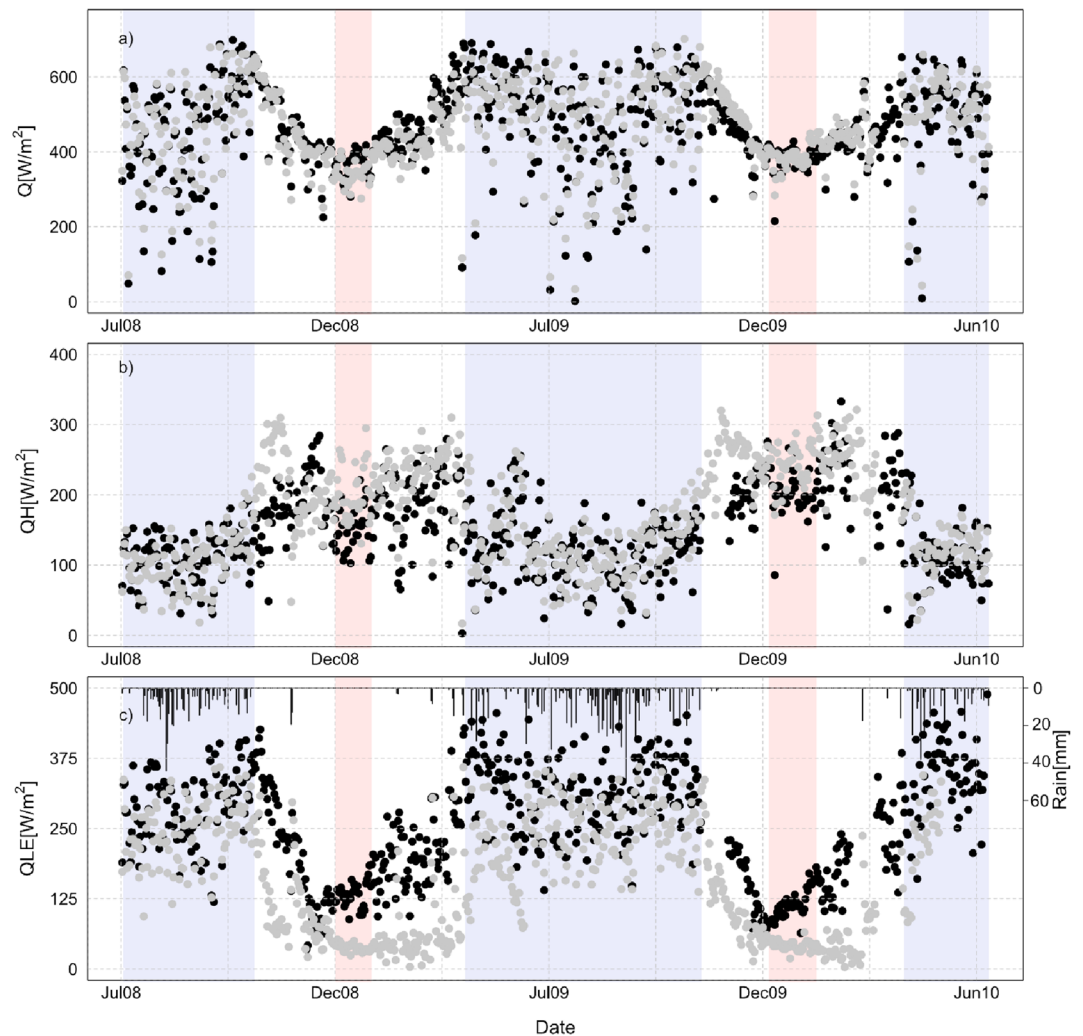
The LAI of herbaceous vegetation at the NA site followed the monsoon regime (Figure 2d). The vegetated period started as soon as the first rains appeared and reached almost  $3 \text{ m}^2 \text{ m}^{-2}$  in the middle of the wet season, corresponding to an herbaceous height of the order of 2.5 m. Then the herbaceous fallow decayed as soon as the rains stopped. The LAI of the woodland was always higher than that of the cultivated area (Figure 2d). The woodland loses its leaves in January–February [Seghieri *et al.*, 2009] leading to a lower LAI during this period. Crop residues are burned on the cultivated site during this same period, but the LAI is nonzero due to isolated shrubs and trees in the SEVIRI pixel. Figure 2 also shows that the interannual variability in meteorological and surface variables is much greater than the intersite differences.

### 3.2. Dynamics of Energy and Water Vapor Fluxes

#### 3.2.1. Midday Averages of Net Radiation

Net radiation ( $Q$ ) followed a similar seasonal pattern at both sites (Figure 3a). From the beginning of the dry season, in 2009, it increased gradually from  $(\sim 352 \pm 28 \text{ W m}^{-2}$  at NA and  $376 \pm 35 \text{ W m}^{-2}$  at BE) to its first maximum





**Figure 3.** Midday averages of: (a) net radiation ( $Q$ ), (b) sensible heat flux ( $Q_H$ ), and (c) water vapor flux ( $Q_{LE}$ ) at Nalohou (gray dots) and Bellefoungou (black dots) during the studied years.

at the beginning of the wet season ( $\sim 580 \text{ W m}^{-2}$ ) and then decreased in the core of the wet season with a large day-to-day variability (coefficient of variation (CV): 24–30%) because of cloud cover variations [Gounou et al., 2012]. The minimum values of  $Q$  observed during dry seasons were caused first by a strong increase of surface temperature which lowered the net longwave radiation and second by the presence of aerosols brought by the Harmattan winds which decreased the net shortwave radiation [Mamadou et al., 2014].

### 3.2.2. Midday Averages of the Sensible Heat Flux

The sensible heat flux ( $Q_H$ ) at both sites (Figure 3b) displayed a strong contrast between dry and wet seasons, with a high day-to-day variability during wet seasons: CV = 34% (NA) and 38% (BE). During dry seasons, high  $Q_H$  values were observed at both sites, resulting from the high surface temperature of the dry soils.  $Q_H$  at BE followed  $Q_H$  at NA closely, except after the wet season when herbaceous fallows were drying out, while the woodland canopy remained green (Figure 2d). In addition, dry vegetation was still tall (roughness length of 0.4–0.5 m) [Mamadou, 2014], which enhanced  $Q_H$  through greater shear. Consequently, together with high available energy (Figure 3a), this induced the highest yearly values of  $Q_H$  at the NA site ( $\sim 330 \text{ W m}^{-2}$ ). At the BE site, however, the highest  $z_0$  was observed later in the year [Mamadou, 2014], linked to the loss of woodland leaves [Seghier et al., 2009] that probably in turn delayed the occurrence of the yearly maximum of  $Q_H$  at this site ( $\sim 290 \text{ W m}^{-2}$ ).

In response to the high soil moisture (Figure 2e) and vegetation growth (Figure 2d) during the wet season,  $Q_H$  strongly declined in association with corresponding increase in  $Q_{LE}$  (Figure 3c) but still remained

significant in the energy balance ( $Q_H/Q_{LE} \sim 0.4$  at both sites). Site to site  $Q_H$  differences vanished with  $Q_H$  values comprised between 50 and  $150 \text{ W m}^{-2}$ .

### 3.2.3. Midday Averages of the Water Vapor Flux

The seasonal variation in water vapor flux ( $Q_{LE}$ , Figure 3c) is in phase with the associated changes in soil moisture and vegetation (Figure 2). The BE site always loses more water vapor than the NA site. Seasonal average differences between sites were about 80 and  $150 \text{ W m}^{-2}$  during wet and dry seasons, respectively (see details in section 4.1). This produced a large difference in annual  $Q_{LE}$  amounts (see further). Midday  $Q_{LE}$  values were the highest as expected during wet seasons on both sites. Although, the NA site was more sensitive than the BE site to rain breaks as observed in May 2009 when  $Q_{LE}$  dropped rapidly for the NA site only. During wet seasons,  $Q_{LE}$  displayed a high day-to-day variability driven by available energy: CV ranged between 25% and 22% at NA and BE, respectively.

During dry seasons,  $Q_{LE}$  values were nonzero at NA ( $\sim 40 \text{ W m}^{-2}$ ), as already observed by Mamadou *et al.* [2014], for the year 2008. Nonzero evapotranspiration rates for cultivated areas in Sudanian climate have also been reported by Bagayoko *et al.* [2007] and farther north in the Sahel by Gash *et al.* [1997], Ramier *et al.* [2009], and Timouk *et al.* [2009]. At the BE site,  $Q_{LE}$  was 3 times higher than at the NA site (Figure 3c), despite the low values of soil moisture. Moreover, soil moisture ( $\theta_{0-1 \text{ m}}$ ) decreased less at the BE site than at NA site (Figure 2e). This suggests that the transpiration activity of woodland in this season was supplied by deeper soil layers (below 1 m depth).

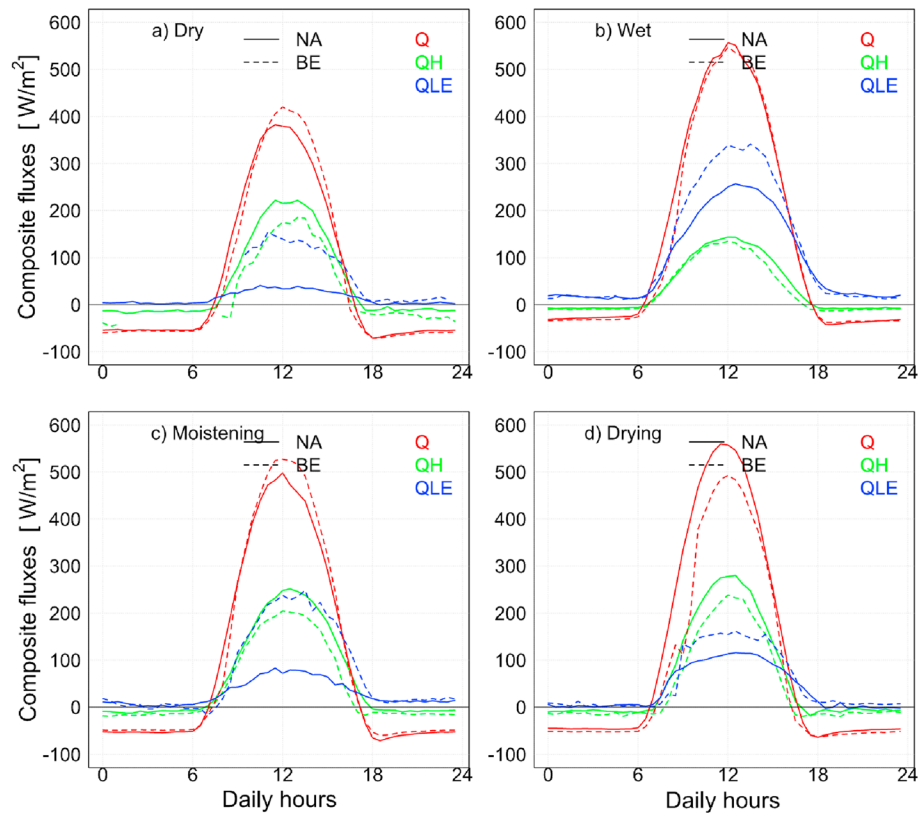
During the moistening period, the difference in evapotranspiration amounts between the two ecosystems was maximum. While  $Q_{LE}$  started to increase at the end of the dry season at the BE site, it remained low and was subjected to isolated rainy events at the NA site. The growth of  $Q_{LE}$  at NA follows more or less the LAI dynamics which started to increase toward the end of the moistening period. Finally, during the drying period, the decrease of  $Q_{LE}$  at the BE site was also delayed compared to the decrease at the NA site. Woodlands evapotranspiration seems to be less sensitive to the lack of rain than the cultivated area.

### 3.2.4. Diurnal Cycles of Exchanges Between Two Ecosystems and the Atmosphere

Figures 4a–4d show seasonal averages of the composite diurnal cycles of net radiation ( $Q$ ), sensible heat ( $Q_H$ ), and water vapor fluxes ( $Q_{LE}$ ) for both sites. The diurnal cycle of  $Q$  at the BE site displays a slight hollow during morning hours caused by branch's shading, especially during the drying period because the measured net radiation at that site corresponded to the understory net radiation. We again observed higher  $Q_{LE}$  at BE than at NA, whatever the season, excepted during night-day and day-night transitions when  $Q_{LE}$  have the same order of magnitude considering measurement uncertainties during these transitions. In the afternoon,  $Q_H$  at NA lasted more than at BE in all seasons. This could be associated with the sustained  $Q_{LE}$  at BE, which probably limits surface temperature through latent heat processes. We finally observed that  $Q_{LE}$  often met and/or exceeded  $Q$  from the afternoon to the evening and remained positive during nighttime, indicating water vapor losses at both sites. Nocturnal evapotranspiration has already been observed at the cultivated site [Mamadou *et al.*, 2014], thus suggesting nocturnal evaporation or an incomplete stomata closure [Caird *et al.*, 2007; Snyder, 2003]. These findings corroborated those obtained by Novick *et al.* [2009] which showed that nocturnal evapotranspiration across a wide range of species in Southeastern US averaged 8–9% of mean daytime evapotranspiration and Awessou *et al.* [2016] that evidenced nocturnal transpiration on *Vitellaria paradoxa* trees of the Bellefougou woodland. On the other hand, nighttime  $Q_H$  was negative in all seasons, meaning that sensible heat was lost from the ecosystem to the atmosphere. This, together with ground heat release, participated in balancing the surface energy loss through longwave radiative emissions, thus allowing ongoing evapotranspiration (Figures 4a–4d).

For the dry season,  $Q_H$  was the main consumer in the energy balance and reached a maximum value of  $215 \pm 46 \text{ W m}^{-2}$  at NA and  $183 \pm 44 \text{ W m}^{-2}$  at BE, at midday.  $Q_H$  at BE was slightly delayed compared with the dynamics of  $Q_H$  at NA, which was not the case for  $Q_{LE}$  that responded to sunrise at both sites. The maximum values of  $Q_{LE}$  at BE occurred in the morning during this season. As soil moisture changed very little from morning to afternoon (less than 0.1 mm) during the dry season, this is probably caused by a temperature limitation of stomatal opening.

During the wet season (Figure 4b),  $Q_{LE}$  became the main consumer of the available energy at both sites.  $Q_H$  cycles were very close despite the vegetation differences. A phase shift of about 1 h can be clearly observed



**Figure 4.** Composite diurnal cycles of energy balance components for the different monsoon periods of the year 2009: (a) dry and (b) wet seasons and (c) moistening and (d) drying periods in the woodland (dotted line) and cultivated (solid line) sites. The net radiation ( $Q$ ), sensible heat ( $Q_H$ ), and water vapor ( $Q_{LE}$ ) fluxes are plotted, respectively, in red, green, black, and blue colors.

between  $Q_H$  and  $Q_{LE}$  at both sites.  $Q_H$  followed the dynamics of  $Q$  more closely than the dynamics of  $Q_{LE}$ , which were delayed in the afternoon.

During the drying period, the soil and the atmosphere become drier.  $Q_H$  became the main consumer of the available energy at both sites with a typical bell-shaped diurnal cycle in phase with that of  $Q$  (Figure 4d), whereas  $Q_{LE}$  was flattened at midday at both sites. The latter is probably associated with the senescence of the vegetation, which becomes less efficient [Gee and Federer, 1972; Seghieri et al., 2009].

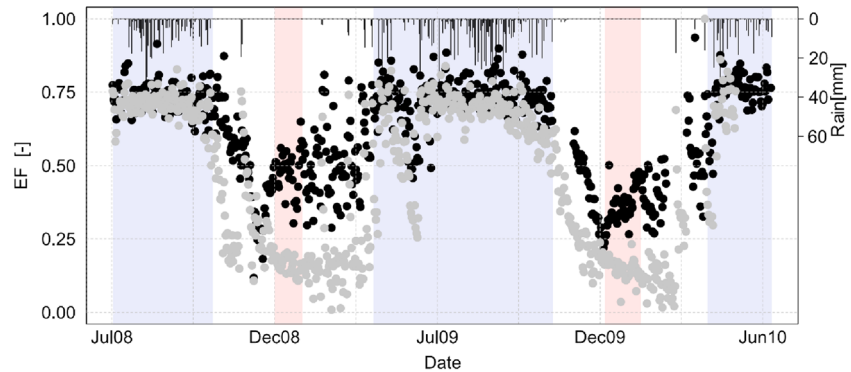
### 3.3. Energy Partitioning and Its Controlling Factors

#### 3.3.1. Environmental and Surface Controls

The evaporative fraction (EF) clearly shows a seasonal cycle at both sites, with some interannual variations that are more distinct in the dry and transitional seasons (Figure 5). It remained rather stable during the wet seasons, especially in the core of the wet season when median EF values equaled  $70 \pm 2.5\%$  at NA and  $75 \pm 0.7\%$  at BE, respectively. This means that 70–75% of the available energy was changed into actual evapotranspiration during the wet season.

The effect of soil water content variations on energy partitioning was further examined via the relationship between EF and the extractable water  $\chi$  (Figures 6a and 6b). A significant positive relationship (Spearman partial correlation  $r = 0.41$  (NA);  $r = 0.36$  (BE); and  $p < 0.0001$ ) was found between EF and  $\chi_{0-30\text{ cm}}$  at both sites (Table 2). EF increased with surface soil water content until a threshold of 50% at NA and 20% at BE, above which  $\chi_{0-30\text{ cm}}$  no longer influenced EF (Figure 6a).

The relationship with the deeper soil layer (30 cm–100 cm) (Figure 6b) is less clear as illustrated by the Spearman partial correlations (Table 2) which are negative at both sites:  $r = -0.28$  (NA) and  $r = -0.26$  (BE).



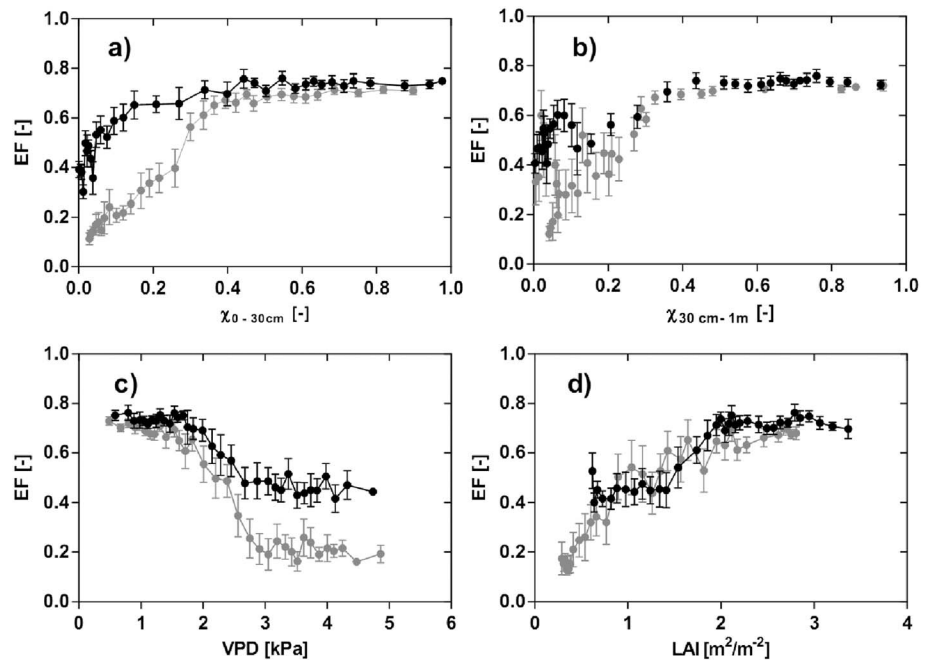
**Figure 5.** Midday evaporative fraction (EF) at Nalohou (gray dots) and Bellefoungou (black dots) during the studied years.

However, as the extractable water contents in the 0–30 cm and 30–100 cm soil layers are intercorrelated ( $r=0.63$ ;  $r=0.46$ ; and  $p < 0.0001$ ), it is not possible to separate their respective influences on EF.

During the wet season, VPD was below 1.7 kPa, (Figure 2c) and seemed to have no visible influence on EF variations at either site (Figure 6c). For high VPD values ( $>3$  kPa), EF also remained more or less stable but with a different level for NA (EF = 20%) and BE (EF = 45%). Between these two VPD values ( $1.7 \text{ kPa} < \text{VPD} < 3 \text{ kPa}$ ), both ecosystems show a linear VPD dependency. The decreasing linear gradient obtained was higher at the NA site ( $-0.38 \text{ kPa}^{-1}$ ) than at the BE site ( $-0.23 \text{ kPa}^{-1}$ ). This likely reflects the sensitivity of AET to VPD during intermediate periods. Both sites show a sensitivity of EF to soil moisture, VPD, and LAI. These variables are intercorrelated. The seasonal contrast drives the energy partitioning and the dynamics of all the parameters.

**3.3.2. Aerodynamic Conductance, Surface Conductance, and Decoupling Coefficient**

The aerodynamic conductance ( $G_a$ ) was nearly constant at NA ( $\sim 100 \text{ mm s}^{-1}$ ), while it showed a strong seasonal variability at BE (Figure 7a). The BE lowest values ( $150 \text{ mm s}^{-1}$ ) were observed during the wet season; during the other seasons, the observed values were systematically higher and reached  $250 \text{ mm s}^{-1}$  in the 2010 dry season. These highest values correspond to the renewing of leaves and a subsequent higher



**Figure 6.** Relationship between midday evaporative fraction (EF) versus: (a) extractable soil water between 0 and 30 cm, (b) extractable soil water between 30 cm and 1 m, (c) the water vapor deficit (VPD), and (d) leaf area index at Nalohou (gray colors) and Bellefoungou (black colors) during the studied years. Each point on the graph is a bin average over 20 measurements. The error bars represent the 95% confidence interval.

**Table 2.** Spearman Partial Correlations Calculated for the Two Studied Years Showing Relations Between Evaporative Fraction (EF) and Some Physical Predictor Variables<sup>a</sup>

CULTURE	EF	VPD	T <sub>air</sub>	χ <sub>0–30 cm</sub>	χ <sub>30 cm–1 m</sub>	LAI
EF	1					
VPD	−0.21 (***)	1				
T <sub>air</sub>	−0.11 (**)	0.65 (***)	1			
χ <sub>0–30 cm</sub>	0.41 (***)	−0.17 (***)	−0.08 (*)	1		
χ <sub>30 cm–1 m</sub>	−0.28 (***)	0.26 (***)	−0.20 (***)	0.46 (***)	1	
LAI	0.23 (***)	−0.36 (***)	0.26 (***)	0.19 (***)	0.48 (***)	1
WOODLAND	EF	VPD	T <sub>air</sub>	χ <sub>0–30 cm</sub>	χ <sub>30 cm–1 m</sub>	LAI
EF	1					
VPD	−0.13 (**)	1				
T <sub>air</sub>	−0.01 (NS)	0.71 (***)	1			
χ <sub>0–30 cm</sub>	0.36 (***)	−0.54 (***)	0.34 (***)	1		
χ <sub>30 cm–1 m</sub>	−0.26 (***)	0.30 (***)	−0.41 (***)	0.63 (***)	1	
LAI	0.22 (***)	−0.18 (***)	0.17 (***)	−0.08 (NS)	0.62 (***)	1

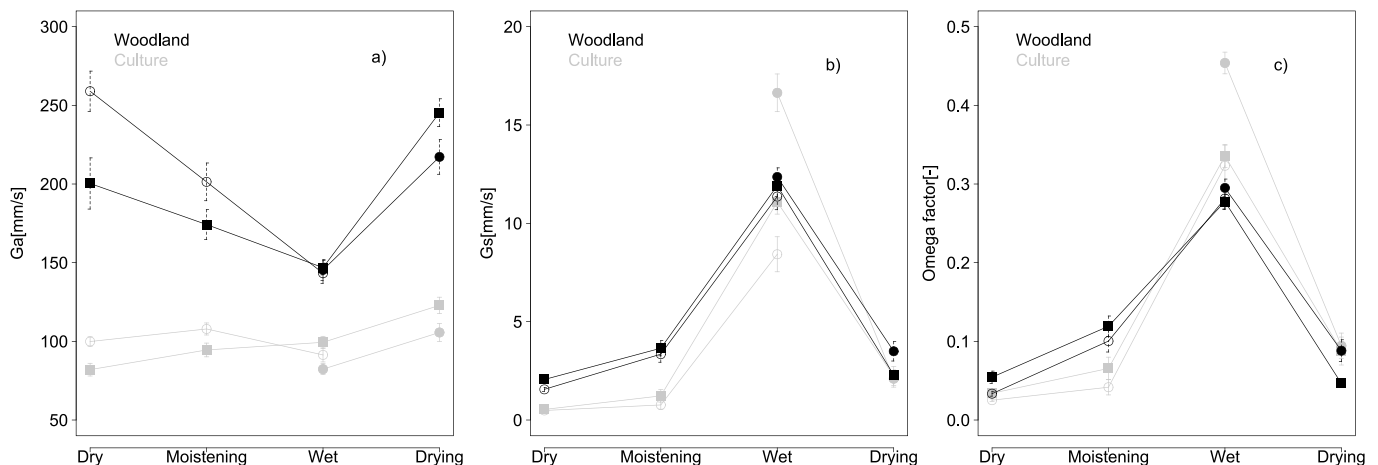
<sup>a</sup>Predictor variables: water vapor deficit (VPD), air temperature (T<sub>air</sub>), extractable soil water in the 0–30 cm layer (χ<sub>0–30 cm</sub>) and 30 cm–1 m layer (χ<sub>30 cm–1 m</sub>) at Nalohou (culture) and Bellefoungou (woodland). The partial correlation (r) and significance level are indicated for each regression; nonsignificant NS (p > 0.05). \*weakly significant for p < 0.01. \*\*moderately significant for p < 0.001. \*\*\*highly significant for p < 0.0001.

roughness length. The highest surface conductance (G<sub>s</sub>) was observed during the wet season (Figure 7b) with mean seasonal maxima values of 11.1 ± 0.6 mm s<sup>−1</sup> (NA) and 12 ± 0.5 mm s<sup>−1</sup> (BE). Figures 7a and 7b also show that the values of G<sub>a</sub> (100–250 mm s<sup>−1</sup>) are far greater than those of G<sub>s</sub> (<20 mm s<sup>−1</sup>), suggesting that the sensitivity of G<sub>s</sub> to G<sub>a</sub> is low as demonstrated by Granier *et al.* [1996]. The aerodynamic resistance (1/G<sub>a</sub>) contribution was indeed always less than 1% of the surface resistance (1/G<sub>s</sub>).

The decoupling coefficient (Ω) (Figure 7c) showed seasonal changes similar to those of G<sub>s</sub>. In the two ecosystems, Ω seasonal averages ranged from 0.05 to 0.45, during the dry and wet seasons respectively. However, in the wet seasons, seasonal maxima values of Ω obtained were slightly lower at BE than at NA, probably because of a higher roughness length at the woodland site (z<sub>0</sub> ~ 1.2 m) than at the cultivated site (z<sub>0</sub> ~ 0.45 m at the end of the wet season) which favored more surface atmosphere coupling through a mixing of the air within the canopy and atmosphere at the woodland site.

### 3.4. Annual Actual Evapotranspiration (AET)

The annual sums of AET for the two studied years are shown in Table 3. For both sites, the measured total annual evapotranspiration varied little over the two studied years. The 12% (resp.15%) increase of annual rainfall from the 2008–2009 to the 2009–2010 year observed at NA (resp. BE) does not seem to have induced



**Figure 7.** Seasonal means values of: (a) aerodynamic conductance (G<sub>a</sub>), (b) surface conductance (G<sub>s</sub>), and (c) decoupling coefficient factor (Ω) at Nalohou (gray) and Bellefoungou (black) during the years 2008 (solid circles), 2009 (squares), and 2010 (empty circles). The error bars represent the 95% confidence interval.

**Table 3.** Seasonal and Interannual Sums of Precipitation (P), Actual Evapotranspiration (AET), Percent of Precipitation That Returns Into the Atmosphere Through Evapotranspiration (AET/P) Over the Woodland and Cultivated Area During the Two Hydrological Years 2008–2009 and 2009–2010

Years	Seasons	Beginning	End	Length (days)	Bellefoungou (Woodland)			Nalohou (Culture)		
					P (mm)	AET (mm ± SE)	AET/P(%)	P (mm)	AET (mm ± SE)	AET/P(%)
2008/2009	Wet	01/07/08	19/10/08	111	880	342 ± 33	39	797	318 ± 21	40
	Drying	20/10/08	26/12/08	68	1	167 ± 16	>100	55	93 ± 10	>100
	Dry	27/12/08	26/01/09	31	0	46 ± 28	>100	0	15 ± 2.4	>100
	Moistening	27/01/09	14/04/09	78	120	203 ± 31	>100	74	75.5 ± 18	>100
	Wet	15/04/09	30/06/09	78	355	289 ± 26	81	289	219 ± 16	76
	Year 1	01/07/08	30/06/09	366	1356	1047 ± 62	77	1215	721 ± 34	59
2009/2010	Wet	01/07/09	31/10/09	123	1119	428 ± 33	38	930	372 ± 27	40
	Drying	01/11/09	26/12/09	56	30	99 ± 35	>100	20	71 ± 25	>100
	Dry	27/12/09	05/02/10	41	0	57 ± 17	>100	0	22.5 ± 3	>100
	Moistening	06/02/10	19/04/10	73	40	178 ± 42	>100	53	71 ± 23	>100
	Wet	20/04/10	30/06/10	73	367	272 ± 45	74	354	199 ± 46	56
	Year 2	01/07/09	30/06/10	366	1557	1034 ± 80	66	1356	736 ± 63	54

a concomitant increase in annual AET (<2%; Table 3). On the other hand, the average annual sum of AET at the woodland site ( $1040 \pm 70$  mm) was 30% higher than that of the cultivated area ( $730 \pm 50$  mm). The ratio of annual AET to annual rainfall was 72% and 57%, respectively, at the woodland and cultivated sites. In the cultivated area, 76% of the annual AET occurs during the wet season ( $554 \pm 55$  mm), while this percentage is only 64% over the woodland ( $665 \pm 68$  mm). The AET amount is distributed more evenly over the year on the woodland site than on the cultivated area.

## 4. Discussion

### 4.1. Ecosystem Differences

Our study indicates that the two ecosystems responded to meteorological (and soil moisture) variations during the dry and intermediate seasons. They were less sensitive to these variables in the wet season but remained mainly driven by the available energy. The intersite differences in the surface energy and water vapor exchanges were much greater than the interannual differences for these two wet years. Both sites had similar meteorological forcing and soil properties; thus, observed surface exchange differences were mainly due to difference in ecophysiological properties.

#### 4.1.1. Wet Seasons

During the wet seasons, the thermal gradient of the surface boundary layer was almost equal [Lohou *et al.*, 2010] in both ecosystems and their exchanges in terms of sensible heat flux were similar (absolute difference lower than  $20 \text{ W m}^{-2}$ ). The difference in  $Q_{LE}$  was greater as  $Q_{LE}$  varied on the average between  $240 \text{ W m}^{-2}$  (NA) and  $320 \text{ W m}^{-2}$  (BE). However, the absolute difference in their evaporative fraction was about 5% (Figure 5). Overall, average values of EF during the wet seasons obtained at BE (0.75) were comparable with reports of 0.73–0.75 obtained over a savanna region in Burkina-Faso [Bagayoko *et al.*, 2007; Brümmer *et al.*, 2008]. At the NA site, EF (0.70) was slightly higher than the 0.67 values reported by [Guyot *et al.*, 2012] from scintillometry measurements over a region with heterogeneous vegetation in northern Benin.

#### 4.1.2. Dry Seasons

During the dry seasons, no rain occurred and the available soil moisture was different for the two vegetation covers, depending on their root system depth. In the cultivated area, the crops grow new roots each year within the first meter of the soil, while the roots of trees in the woodland continue to grow and can reach deeper soil over the years. This factor led to both higher actual evapotranspiration (Figure 3c) and transpiration rates ( $G_s \sim 3 \text{ mm/s}$ ; Figure 7b) at the woodland site. Similar results were found by Juárez *et al.* [2007] and Vourlitis *et al.* [2002] over the Amazonian forest where the dry season lasts more than 4 months without any rainfall. These high transpiration rates are due to vegetation activity which even leads to leaf renewal during this dry season in the region [Seghieri *et al.*, 2009]. It is also associated with high physiological activity as shown in the sap flow analysis [Awessou *et al.*, 2016] as well as in the  $\text{CO}_2$  fluxes [Ago *et al.*, 2016, 2015, 2014] in the studied area. This phenomenon was also observed farther north above a savanna in Ghana by Quansah *et al.* [2015].

#### 4.1.3. Comparison With Other Tropical Ecosystems

The role of deep roots in the transpiration process, especially above forests, has been pointed out in other climatic contexts [Collins and Bras, 2007; Feddes et al., 2001; Nepstad et al., 1994]. In the Sudanian climate, the soil moisture in the 0–1 m layer was low and did not vary too much during dry seasons (Figure 2e), which indicates that the trees extract water from deeper soil layers. Since the maximum rooting depth is about  $15 \pm 5.4$  m for tropical grassland/savanna [Canadell et al., 1996], this probably favors the extraction of water from deep layers, since a depletion of permanent groundwater was already observed during the dry season in the region [Séguis et al., 2011; Vouillamoz et al., 2015]. The ability of trees to extract available water from deep soil layers allows them to survive in the dry season. Consequently, the major differences in AET between the two ecosystems are observed during the moistening and the drying periods (Table 3).

On the average, annual actual evapotranspiration represents 57% of the rainfall at the cultivated site and 72% at the woodland site (Table 3). These percentages are in the same range but slightly higher than those obtained over Sahelian sites (Niger—Lat: 13.65°N–600 mm yr<sup>-1</sup> of rain). Using the eddy covariance technique, Ramier et al. [2009] found, on the average for a two years period, that AET was 65% (resp. 45%) of the annual rain amount for a fallow (resp. a millet crop) site. In another tropical region, a 7 year average of 52% was found by Vourlitis et al. [2015], using the eddy covariance technique above a semideciduous forest in the rainforest-savanna of northern Mato Grosso, Brazil. All these observations are probably a lower bound for the total amount of AET because of underestimation of turbulent fluxes associated with the eddy-covariance technique [Foken et al., 2011]. This underestimation was already pointed out by Velluet et al. [2014] whose simulations showed a 4 year AET average of 82–85% of the total amount of rain for the Niger rain-fed millet crop and fallow bush, whereas Ramier et al. [2009] measured 45%–65% over the same sites. However, differences between sites subjected to the same atmospheric forcing can obviously be compared and analyzed in terms of land cover change. The difference of ~300 mm between the two sites is probably a reliable order of magnitude.

#### 4.2. Implication for Land Surface Feedbacks and Water Balance

The lower albedo and surface temperature of woodlands [Bonan, 2008] increases net radiation, thereby favoring water vapor flux losses in the energy budget partition (Figure 3c). The conversion of woodland into cultivated areas would thus probably induce some changes in land surface feedbacks, returning less water vapor to the atmosphere (Table 3) and therefore less latent heat energy. At the same time, on an annual basis, cumulative sensible heat fluxes are close from one site to another. Therefore, in terms of energy transfer, these observations suggest that land conversion from woodland to cultivated areas in this Sudanian region would probably not participate greatly to the heating of the atmosphere through sensible heat fluxes but would reduce the latent (and hence the water vapor) fluxes to the atmosphere, specifically during the moistening period (Figure 3c). This atmosphere moistening process has been identified [Lothon et al., 2008] as a major contributing factor in the onset of the rainy season. A reduction in the moistening of the atmosphere by the surface should therefore impact the onset of the rainy season and probably the monsoon dynamics over the region [Koster et al., 2004].

Land conversion from woodlands to cropping areas should also affect the hydrologic cycle, especially the groundwater dynamics. For instance, land clearing would be expected to produce a rise in the groundwater level as deep water uptake is reduced. Such processes have been reported in other regions [Kelly et al., 2016; Krishnaswamy et al., 2012; Moore et al., 2012; Peña-Arancibia et al., 2012; Zhang et al., 2012]. However, despite drastic deforestation since in Benin 1973 (from 70% to 25% of the total surface), the upper Oueme river discharge (a 15,000 km<sup>2</sup> catchment in which the Donga catchment is embedded) has not shown any great change [Lelay and Galle, 2005; Peugeot et al., 2011]. As shown by Hector et al. [2015] and Séguis et al. [2011], river flows over recent years in this area mainly originate from perched, seasonal water tables and the contribution of deep (10–20 m) groundwater is negligible. At the hillslope scale, Richard et al. [2013] conducted a virtual experiment that showed that the riparian forest plays a major role in disconnecting the water table from the river flow. Annual aquifer recharge in northern Benin has been quantified to be between 100 and 400 mm depending on the ground lithology [Vouillamoz et al., 2015]. The AET decrease due to deforestation lies within this range and we have shown that it is partially drawn from deep ground layers. These first AET observations in northern Benin therefore gave rise to many questions concerning the water budget: How aquifer recharge would be impacted by land cover changes? Would ground storage capacity be sufficient to store the excess of the water budget? What is the effect of rainfall variability and land cover changes on river discharges and water table level? To answer such questions, longer series of water budget observations will

be required along with hydrological models that can simulate vertical and horizontal water transfers from rain to the water table and rivers, taking into account surface and vegetation processes. Changes in the continental hydrological cycle could also in turn impact feedbacks to the atmosphere related to increased accessible water close to the surface. Once again, models including both the continental and atmospheric hydrological cycle could help us to answer such questions [Maxwell *et al.*, 2010].

## 5. Conclusion

This study, using 2 years of eddy covariance data, is the first to compare the dynamics of water vapor and sensible heat fluxes of a woodland and a cultivated area in Sudanian West Africa. To our knowledge, it represents the longest continuous field scale data set of turbulent fluxes of these ecosystems published so far for the Sudanian climate. At seasonal scale, we have shown that soil moisture is the main controlling factor of the net radiation partitioning into sensible and water vapor fluxes. The seasonality of the sensible heat flux has been clearly demonstrated for both sites. Depending on the season, the seasonal cumulative water vapor loss of the woodland differed from that of the cultivated area by a factor of 1.2 (wet season) up to 3 (dry season). Globally, the AET amount is distributed more evenly over the year on the woodland site than on the cultivated area. The average annual AET measured during the investigated years represented 57% and 72% of the annual rainfall over the cultivated area and the woodland, respectively. The differences in the water vapor fluxes of the two ecosystems (30%) were much higher than the interannual variability (<2%). They could impact the hydrological cycle, but the available time series are not long enough to draw firm conclusions. Moreover, the studied years were characterized by wet atmospheric conditions compared to the 50 year local mean precipitation. For this reason, we have not been able to clarify the effects of woodland and cultivated land cover under water stress conditions. More years of measurements including dry weather conditions are thus necessary to consolidate these findings.

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