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Simulation of Surface Fluxes in Two Distinct Environments along a Topographic Gradient in a Central Amazonian Forest using the INtegrated LAND Surface Model

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Abstract

The Integrated Land Surface model (INLAND) land surface model, in offline mode, was adjusted and forced with prescribed climate to represent two contrasting environments along a topographic gradient in a central Amazon *Terra Firme* forest, which is distinguished by well-drained, flat plateaus and poorly drained, broad river valleys. To correctly simulate the valley area, a lumped unconfined aquifer model was included in the INLAND model to represent the water table dynamics and results show reasonable agreement with observations. Field data from both areas are used to evaluate the model simulations of energy, water and carbon fluxes. The model is able to characterize with good accuracy the main differences that appear in the seasonal energy and carbon partitioning of plateau and valley fluxes, which are related to features of the vegetation associated with soils and topography. The simulated latent heat flux (LE) and net ecosystem exchange of carbon (NEE), for example, are higher on the plateau area while at the bottom of the valley the sensible heat flux (H) is noticeably higher than at the plateau, in agreement with observed data. Differences in simulated hydrological fluxes are also linked to the topography, showing a higher surface runoff (R) and lower evapotranspiration (ET) in the valley area. The different behavior of the fluxes on both annual

and diurnal time scales confirms the benefit of a tiling mechanism in the presence of large contrast and the importance to incorporate subgrid-scale variability by including relief attributes of topography, soil and vegetation to better representing *Terra Firme* forests in land surface models.

Key words: Landscape heterogeneity; plateau; valley; land surface model; water balance; carbon balance; central Amazon; *Terra Firme* forest soil and vegetation parameters.

1. Introduction

The Amazon Rainforest, which has an area of more than 6.3 million km^2, contains approximately 50% of the world's tropical forests (Pan et al., 2011). The region is a mosaic of landscapes which are traditionally divided into two major forest types: inundated (over alluvial terrain) and non-inundated upland (*Terra Firme*) forests (Prance et al., 1979). *Terra Firme* forests are found in well-drained areas with an elevation lower than 100 m; these forests cover a high percentage of the Amazon. In the central region, for example, *Terra Firme* forests represent 80% to 90% of the area (Ayres, 1995, Hess et al., 2003) or perhaps more (Anderson et al., 2009). Furthermore, these forests contain approximately 74% of the terrestrial biomass in the Brazilian Amazon basin (Vieira et al., 2004), including complex strata of emergent trees, canopy trees, understory trees, understory shrubs, saplings, seedlings, herbs and ferns. When viewed on a large scale, Amazonian *Terra Firme* forests appear to form a homogeneous plain that is structurally uniform. However, when analyzed at a smaller scale (spatial scales on the order of tens of meters to a few kilometers), there are diverse environments that are determined by topography, soil type and soil moisture regime that are often masked by dense forest cover (Nobre et al., 2011). In the central Amazon, a relatively small area of *Terra Firme* forest usually comprises the topographic gradient of the plateau, slope and valley environments (Chauvel et al*., 1987;* Araújo et al., 2010; Hodnett et al., 1997, Nobre et al., 2011; Rennó et al., 2008). These environments are characterized by a high variability in species composition, which is at odds with the homogeneous appearance found in satellite images. The plateaus are approximately 50 to 90 m higher than the valley bottoms, with moderately steep slopes between them. The valley bottom areas are swampy, with pools of water and small streams which are locally known as *Igarapés* (Chauvel et al., 1987). Groundwater from beneath the plateau and slopes continuously flows to the valley where it maintains the water table close to the surface, even in most dry seasons (Hodnett et al., 1997). Using this elevation range as a threshold, Anderson et al. (2009) found that 17% of the *Terra Firme* forests in this landscape can be considered a valley vegetation type. According to Nobre et al. (2011), in the central Amazon, which is representative of other extensive areas in Amazonia, the valley forest environment covers 26.2% of the area, whereas the slope and plateau forests occupy 30.7% and 43.1%, respectively. Miguez-Macho and Fan (2012) found, using a numerical model, that the water table exists between the surface and a depth of 2 m across 20 - 40% of the area of the Amazon throughout the year.

The local topographic heterogeneity is an important factor that governs the diversity of elements comprising the landscape in *Terra Firme* forests (Liberman et al., 1985; Takyu et al., 2002). This topographic heterogeneity exerts a significant influence on soil composition and the distribution of vegetation (Pelissier et al., 2001). The type of vegetation in each topographic zone is associated with differences in leaf area index, which influences the radiation balance and the gas and energy exchange between the plant environment and the atmosphere (Araújo, 2009; Leitão 1994; Oliveira, 2010). Thus, the topographical heterogeneity is also associated with carbon exchange changes between the forest and the atmosphere (Araujo et al., 2008). The topography also affects the water content and the water dynamics of the soil (Tomasella et al., 2008; Ferreira et al. 2002) and plays a decisive role in the establishment of areas influencing water table variations (Cuartas et al., 2012; Nobre et al., 2011). These factors affect the infiltration process and runoff as well as the amount of water available in the soil for extraction by plant roots, which is associated with the transpiration process (Tomasella et al., 2008). The topographic effects on hydrology, soils and vegetation described here are certainly not due the topography *per se*, but to environmental conditions defined by the topography. Although these fluxes can be quantified through land surface models (LSMs) or soil-vegetation-atmosphere-transfer models (SVATs), their coarse grid resolutions and simplified parameterizations of subgrid-scale heterogeneity induce errors in the modeled mean fluxes due to the strong nonlinearity and fine-scale heterogeneity of land surface processes. There are still no studies that explicitly incorporate subgrid-scale variability by including relief attributes (topography), soil and vegetation to better represent *Terra Firme* forests in these models. Developing models with this perspective will allow progress in the representation of the Amazon basin at the large scale within integrated Earth System models. However, to better understand its role in the global climate and how each topographical type is associated with the effects of vegetation, soil and soil moisture dynamics on the regional water budget, a better understanding of smaller-scale features is needed. The main objective of the present work was to determine and characterize the necessary components (including soil, vegetation and hydraulic) of the Integrated Land Surface (INLAND) model to simulate the plateau and valley environments in a primary forest in the central Amazon. To assess the quality of the model, we evaluated the surface fluxes over a plateau and valley and compared them against *in situ* observations. We hypothesized that the model would simulate the differences in fluxes between the plateaus and valleys because the energy exchange dynamics of these ecosystems are different due to the large diversity in their surface characteristics.

2. Materials and methods

a. Study area description

The study area is an instrumented hydrological catchment of 6.37 km^2 (54° 58′W, 2° 51′S) that is located in the Cuieiras Biological Reserve, also known as ZF2, which belongs to the National Institute for Amazon Research (INPA). The site is 80 km northwest of Manaus in the central Brazilian Amazon rainforest (Figure 1). This site is typical of the topography of much of central Amazonia and the vegetation is typical of undisturbed primary tropical forests or dense *Terra Firme* forests (Ferreira et al., 2005). The climate in central Amazonia is tropical rainforest climate (Af) according to the Koppen classification (Alvares et al., 2014), with monthly average temperatures between 25 °C in July and 27 °C in November and an average relative humidity that exceeds 80% (Leopoldo et al., 1987). The mean annual rainfall is approximately 2000 mm, although it can vary from 1400 to 2800 mm. The rainy season is from November to May, and a dry (or less wet) season occurs from June to October. Approximately 73% of total precipitation, falls during short, heavy rainfall events (Leopoldo et al., 1987).

The site area is a headwater catchment consisting of plateau areas that vary in height between 90 and 110 m above sea level (asl) and are strongly incised. The slopes are steep, with a concave form leading to rather broad, flat valley bottoms (much narrower at the head) at 40-60 m asl. Over the 6.37 km^2 catchment area, the valley bottoms occupy 43% of the surface, and the slope and plateau areas occupy 26% and 31%, respectively (Nobre et al., 2011). The vegetation cover on the plateau and slope areas is composed of tall and dense *Terra Firme* (non-flooding) tropical forest, with canopy heights varying between 30 to 44 m. On the valley

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floor, the vegetation cover is less dense, with canopy heights from 15 to 25 m (Oliveira et al., 2002). The soil composition is strongly controlled by the topography and can be classified into three dominant types along the topographic gradient: clayey Oxisols on the plateaus (Yellow Latosols in the Brazilian system), transitioning to less clayey Ultisols on the slopes (Podzol in the Brazilian system) and ending with the sandy Spodosols on the valley bottoms (Chauvel et al., 1987). These soils are typically acidic and very low in nutrients such as phosphorus, calcium, and potassium (Bravard and Righi, 1989). In the valley bottoms, the water table remains near the surface for most of the year, often up to 100 m from the stream, and these areas often contain swampy pools. A more detailed description of the site can be found in Araújo et al. (2002), Waterloo et al. (2006), Cuartas et al. (2007), Luizão et al. (2004) and Chambers et al. (2004).

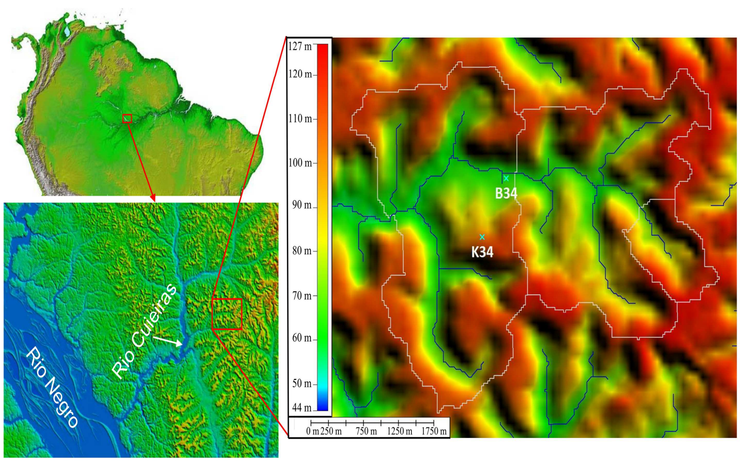

Figure 1. Location of the study area in in the Cuieiras Biological Reserve, Manaus, Amazonas, Brazil.
Source: Adapted from Cuartas et al. (2012)

b. Site measurements

The datasets used in this study include meteorological and hydrological data which were provided by the Large-Scale Biosphere-Atmosphere (LBA) Experiment in the Amazon (Figure 2). The meteorological data were obtained from two micrometeorological eddy flux towers located 600 m apart: one on the plateau and the other on the valley bottom, with an elevation difference of approximately 60 m. The temporal resolution of the data used in this study is 60 minutes. The plateau tower (known as K34, 2° 36’32.67”S, 60° 12’33.48”W, 54 m asl) provided the data for this study from January 2000 to December 2011, including air temperature (°C), precipitation (mm), downward shortwave and longwave radiation (W m^{-2}), wind speed (m s^{-1}), relative humidity, carbon dioxide (CO_2, μ mol m^{-2} s^{-1}), and latent (LE, W m^{-2}) and sensible (H, W m^{-2}) heat fluxes. The valley bottom tower (known as B34, 2° 36'09.8"S, 60° 12'44.5"W, 42 m asl) provided data from January 2006 to December 2006, including CO_2 and LE and H fluxes. All measurements were made from the top of the towers. More details about the K34 and B34 tower instrumentation and the measured variables can be found in Araújo et al. (2002) and Araújo (2009).

The hydrological data obtained was the soil water content in the plateau and stream discharge and water table level in the valley. The soil water content (m^3 m^{-3}) was obtained using two instruments: neutron probe (Didcot Instrument Co., Burwell, Cambridge, UK) and time domain reflectometry - TDR (Campbell Scientific Inc. CS615, Logan, Utah, USA), which are both located on the plateau, approximately 30 m from the K34 meteorological tower (Figure 2). Neutron probe measurements were obtained from 3 access tubes (T1, T2 and T3) installed from the surface to a depth of 4.8 m, with weekly or biweekly measurements from December 2001 to December 2006 (Cuartas et al., 2012). The soil water content was also obtained from TDR sensors between depths of 0.8 to 8.0 m, which were installed along the walls of a lined 15 m shaft, with continuous data recording at hourly intervals between January 2003 and February 2006 (Broedel et al., *2017*). Furthermore, water table depth (m) and stream discharge (m^3 s^{-1}) data were used for the study area during the period from July 2002 to October 2006. The stream discharge (m^3 s^{-1}) was measured using the ultrasonic Doppler technique (Starflow model 6526, UNIDATA, O' connor, Fremantle, Australia) in the valley area. The water level and average and maximum discharge velocity data necessary to obtain the stream discharge were logged every 30 minutes and downloaded from a data logger on a weekly basis. Details on the derivation of the stream discharge data can be found in Waterloo et al., (2006). The water table depth was measured weekly with 7 piezometers spread over the valley area, consisting of perforated Polyvinyl Chloride (PVC) tubes with a diameter of 5 cm and filters (Eijkelkamp Agrisearch Equipment, Nijverheidsstraat, Giesbeek, The Netherlands) (Tomasella et al., 2008).

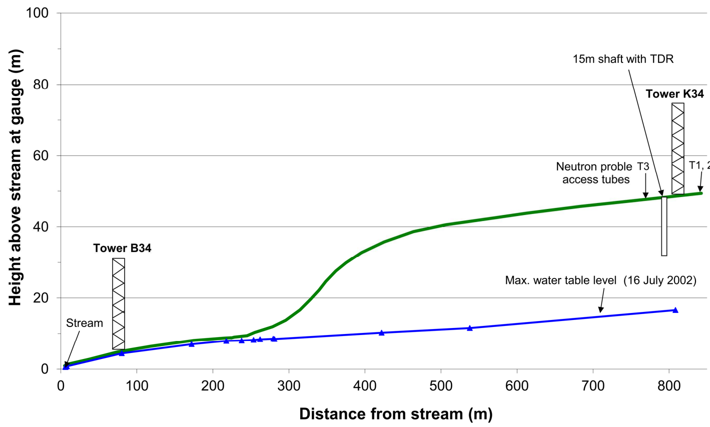

Figure 2. Topographical gradient in a tropical rain forest in the central Amazon and instrumental data. Blue triangle symbols show the height of the water table at the positions of boreholes for monitoring the water table.

d. *The INLAND model*

Version 1.0 of the INtegrated LAND surface model (INLAND) was used in this study (Tourigny, 2014). The INLAND model is a Brazilian development based on the Integrated Biosphere Simulator (IBIS) model (Foley et al., 1996; Kucharik et al., 2000), including all of its main features, using a modular, physically consistent framework to perform integrated simulations of water, energy, and carbon fluxes, but with improvements to simulate particular tropical processes. Here we discuss the components of INLAND that are most relevant to the focus of this paper. The model consists of four component modules organized with respect to their characteristic temporal scales: a land surface module (minutes to hours), a vegetation phenology module (days to weeks), and carbon balance and vegetation dynamics modules, which both have a temporal scale of years. The land surface module of INLAND is taken from the second-generation LSX model (Pollard and Thompson, 1995) and includes 6 soil layers (with varying thicknesses) and an upper (trees) and lower (shrubs and grasses) canopy. Plants are represented by 12 plant functional types (PFTs), each with distinct carbon pools for leaves, stems and roots. The Amazon basin in INLAND model is predominantly represented by the *tropical broadleaf evergreen* tree PFT. The soil module in INLAND simulates soil temperature, water content, and ice content (when required) in each of the 6 soil layers and solves the θ-based form of the Richards equation, where the soil moisture change in time and space is a function of soil water retention curve, soil hydraulic conductivity, upper and lower boundary conditions and plant water uptake. The plant root-water uptake (represented by a sink term in the macroscopic Richard's equation) is a function of atmospheric demand, soil

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physical properties, root distribution, and soil moisture profile (Kucharik et al., 2000).The drainage from the bottom soil layer is modeled assuming gravity drainage and neglects interactions with groundwater aquifers.

Energy balance in the INLAND model is separately calculated for each surface type. Under steady-state conditions, the balance between incoming and outgoing radiation or net radiation at the surface (Rn) must equal the sum of sensible (H), latent (LE) and ground (G) heat fluxes, heat storage within the vegetation canopy (S) and the sum of additional energy sources and sinks (Q). Both S and Q are frequently neglected because of their small magnitudes. Net radiant energy is partitioned between latent, sensible and ground heat fluxes following a Penman-Monteith approach (for details see Monteith, 1965). The model calculates the canopy-level surface carbon balance through the flow of carbon between the atmosphere and plants, which is also known as the net ecosystem exchange (NEE). Negative NEE values correspond to the net carbon uptake by the land surface. The NEE is computed as the difference between heterotrophic respiration (Rh) and net primary productivity (NPP). NPP refers to the carbon that remains stored in plants after taking up CO_2 from the atmosphere during gross primary productivity (GPP) and that has partially respired back through autotrophic respiration (Ra). Leaf-level photosynthesis in the model is determined by the formulations of Farquhar (Farquhar et al., 1980). In these formulations, the photosynthesis rates are a function of absorbed light, leaf temperature, CO_2 concentration within the leaf, and the Rubisco enzyme capacity of photosynthesis. The stomatal conductance is dependent on the photosynthetic rate, CO_2 concentration, and water vapor concentration (Foley et al. 1996). The current version of the model uses a single-leaf photosynthesis approach, and the coupling between photosynthesis and canopy conductance is based on vapor pressure deficit (Leuning, 1995).

e. Development of an unconfined aquifer model

The INLAND model does not explicitly represent water table dynamics. Instead, the lower boundary condition is allowed to vary from 100% free drainage to zero flux (based on an empirical coefficient ranging from 0 to 1). This means that the model can be applied to a plateau environment, although it cannot correctly simulate a valley environment where the water table remains at or very close to the soil surface for much of the year. Consequently, we incorporated into the INLAND model a lumped unconfined aquifer model developed by Yeh and Eltahir (2005 a, b), in which the water table is interactively coupled to the soil column through the soil drainage (groundwater recharge) fluxes. All processes within INLAND, except for those computing the soil moisture, were preserved with the original IBIS equations. The lumped water balance equation for an unconfined groundwater aquifer can be written as follows, according to Yeh and Eltahir (2005 a):

$$S_y \frac{dH}{dt} = I_{gw} - Q_{gw} \qquad (1)$$

where Sy (dimensionless) is the specific yield of the unconfined aquifer, H (m) is the water table level above the datum, Igw (mm day^{-1}) is the groundwater recharge flux, which is the flux at the interface between the unsaturated and saturated zone, i.e., the water table, and Qgw (mm day^{-1}) is the groundwater discharge to streams (i.e., groundwater runoff). For the sandy soils typical of the valleys in the central Amazon rainforests, the value of Sy was specified as 0.265, which is based on specific yield data compiled by Johnson (1967). Yeh and Eltahir (2005a) identified a strong nonlinear relationship between the baseflow discharge and water table depth. A regression analysis was performed on 5 years of baseflow discharge and water level data from our study site; this analysis indicated a similarly strong relationship (Eq. 4), with a correlation coefficient of 0.85 (Figure 3). The baseflow was separated from the daily discharge using a digital recursive filter technique (Lyne and Hollick, 1979). This method has been widely used for the continuous partitioning of streamflow discharge between surface runoff and baseflow because it is fast, efficient, reproducible and objective (e.g., Furey and Gupta, 2001; Nathan and McMahon, 1990 and Lo et al., 2008). Yeh and Eltahir (2005 a) showed that a digital recursive filter can have a high coefficient of determination (i.e., R^2) of 0.84, while Arnold and Allen (1999) found a coefficient of determination of 0.86.

$$Q_{gw} = \frac{0.5156}{(D_{gw})^{0.796}} - 1.4 \qquad (2)$$

The groundwater model represented by Eq. (3) and Eq. (4) was interactively coupled with the soil model in INLAND such that the total length of the active unsaturated soil column varied in response to the water table depth fluctuations by keeping the number of unsaturated layers variable.

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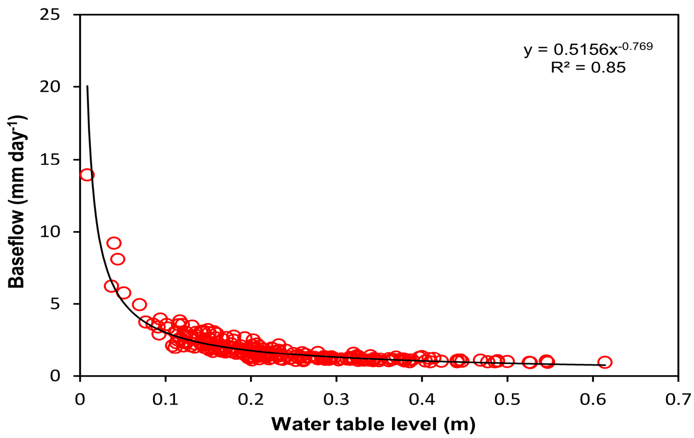

Figure 3. Regression analysis, including water table depth versus monthly baseflow, from July 2002 to October 2006 in the valley study area.

The INLAND model includes a non-linear root water extraction scheme to describe the impact of soil water stress (Folley et al., 1996), in which potential transpiration is first distributed over the rooted zone and then reduced to actual root-water-uptake by a soil water stress reduction function. However, after incorporating the unconfined aquifer model into INLAND to represent the water table dynamics of the valley site, it was also important to represent the stress due to saturated (or waterlogged) conditions, which lead to oxygen deficiency (hypoxia) in the soil. These conditions cannot be ignored because they influence plant survival, growth and functioning at the valley sites (Pezeshki and DeLaune, 2012). We used the linear function presented by Feddes et al. (1978) to describe the soil water uptake reduction caused by root oxygen deficiency, which is added to Folley function (Figure 4). Under optimal moisture conditions, the maximum possible root water extraction rate integrated over the rooting depth is equal to the potential transpiration rate, whereas, under non-optimal conditions (i.e. when the soil is either too dry or too wet) the root water extraction rate may be reduced, causing a reduction in transpiration (Figure 4). For saturated conditions, the reduction of root water uptake occurs between 0.83 – 1.0 degree of soil saturation, which leads to a reduction of 0-17% in the plant transpiration.

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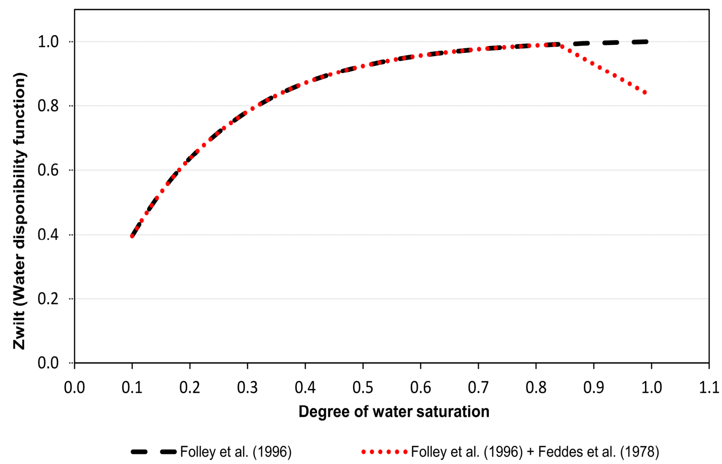

Figure 4. The general shape of root water extraction from Folley et al. (1996) and Feddes et al. (1978) for dry and wet conditions, respectively.

f. Experimental design

All of the simulations reported here in were run with the single-point off-line version of the model (0D, uncoupled from a Global Circulation Model or GCM) with the CO_2 concentration set to a constant value of 400 parts per million (ppm). INLAND was forced using the observed hourly meteorological data collected at the K34 tower for 12 full years, from January 1, 2000, to December 31, 2011. For the plateau simulations, the top-to-bottom thicknesses of the 6 soil layers in INLAND were set to 0.20, 0.30, 0.50, 1.0, 2.0 and 4.0 m, resulting in a total depth of 8 m. For the valley, the total profile depth was set to 4 m, with layer thicknesses of 0.10, 0.20, 0.30, 0.40, 1.0 and 2.0 m from the top to the bottom. We conducted two sets of simulations, with a time step of 60 minutes, for each location (plateau and valley); in the first set of experiments, to avoid additional complexity, the vegetation was set to fixed, or "static", in the INLAND vegetation dynamics module. Thus, no change in the stand structure was assumed to occur during the simulation period. For simplicity, we used the same forcing data for the spin-up period, which is a preliminary simulation to equilibrate the model parameters, such as soil temperature and soil moisture. The spin-up in our simulations was done for duration of 60 years, initialized on January 1, 1940 and was discarded in the analysis. In the second set of experiments, the dynamic vegetation routine was employed in order to evaluate changes in vegetation cover (biomass stocks) and carbon fluxes (productivity) between the plateau and valley after 100 years of integration (from

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January 1, 1900 to December 31, 1999). This approach consists of two different sets of simulations for both the plateau and valley environments: a dynamic vegetation run, which starts from initial vegetation similar to the current tropical forest biome or *dynamic vegetation 1 (DV1),* and a "cold start" run that starts from bare soil and lets the model build the vegetation cover according to climate conditions or *dynamic vegetation 2 (DV2)*. Plant growth, which results from the assimilation of C through photosynthesis, contributes to the formation of a canopy and is moderated by competition-related mortality. Carbon pools in live and dead biomass and in the soil are continuously updated and provide a "memory" of the state of the system over a series of years to decades (Folley et al. 2000).

In the original 0D version of INLAND, the values for the soil hydraulic parameters are obtained from soil texture-based look-up tables commonly used in LSMs, based on pedo-transfer functions (PTFs), such as those of Clapp and Hornberger (1978) and Rawls et al. (1982). However, Hodnett and Tomasella (2002) showed that these PTFs, which were originally derived using data from temperate region soils, do not accurately predict the properties of tropical soils, particularly Oxisols. Although Oxisols have a high clay content, they have low water availability and highly saturated conductivity; these characteristics differ from temperate clay soils. Therefore, these look-up tables were replaced with hydraulic properties for Oxisols taken from Tomasella and Hodnett (1996), Ferreira et al. (2002) and Marques (2009). On the other hand, hydraulic properties for sandy soils located in valley area was taken from Campbell and Norman (1998) and Verhoef and Egea (2014). Furthermore, different soil hydraulic parameters were used for each layer of the soil profile. The performance of the INLAND model was evaluated by comparing the simulations against the observational data from both the plateau and valley areas, including LE, H, Rn, NEE, soil moisture and water table level; three indices of error statistics were used (Ambrose and Roesch, 1982): bias, coefficient of determination (R^2), and the Root Mean Square Error (RMSE).

3. Results and discussion

a. Representation of water table dynamics in the valley

After identifying the optimal vegetation and soil parameters set of plateau and valley forest (Table 1), and incorporating the unconfined aquifer model into INLAND to correct represent the valley area, we compared the water table depth (WTD) simulated by model with observations, from 2002 to 2006.

Table 1. Vegetation and soil parameters used in the INLAND simulations for the Plateau and Valley. The initial values refer to the parameters used by Imbuzeiro (2005).

Parameter	Description	Unit	Initial	Plateau	Valley
Vegetation					
fl	Fraction of overall area covered - lower canopy	-	1.00	1.00	0.90
fu	Fraction of overall area covered - upper canopy	-	0.50	0.50	0.60
dispuhf	Zero-plane displacement height - upper canopy	m	0.70	0.70	0.60
rhoeveg_NIR	Reflectance of an avg leaf/stem near infrared - upper canopy	-	0.400	0.310	0.260
plaievgr	Initial total LAI - upper canopy	$m^2\ m^{-2}$	5.00	4.60	4.30
ztop	Canopy height	m	30	35	18
beta1	Parameter for Jackson rooting profile - lower canopy	-	0.950	0.950	0.940
beta2	Parameter for Jackson rooting profile - upper canopy	-	0.975	0.970	0.965
V_{max}	Maximum Rubisco capacity of the top canopy at 15°C	mol $CO_2\ m^{-2}\ s^{-1}$	$80x10^{-6}$	$80x10^{-6}$	$70x10^{-6}$
coefm	m coefficient for stomatal conductance relationship	-	8.0	9.0	7.5
tauleaf	Foliar biomass turnover time constant	years	1.01	1.02	1.01
tauwood0	Wood biomass turnover time constant	years	25.0	45.0	35.0
tempvm	Stress maximum coefficient of V_{max}	-	3500	6000	4500
tdripu	Decay time of liquid intercepted by leaves - upper canopy	s	7200	16200	10800
tdrips	Decay time of liquid intercepted by stems - upper canopy	s	7200	16200	10800
tdripl	Decay time of liquid intercepted by leaves and stem - lower canopy	s	7200	16200	10800
Soil					
swilt	Wilting point	$m^3\ m^{-3}$	0.272	0.370	0.030
suction	Air entry potential	m_H_2O	0.370	0.100	0.050
sfield	Fiel capacity	$m^3\ m^{-3}$	0.369	0.430	0.090
poros	Porosity	$m^3\ m^{-3}$	0.475	0.48; 0.52; 0.52; 0.575; 0.59; 0.595	0.40; 0.41; 0.42; 0.43; 0.44; 0.45
bex	Campbell 'b' exponent	-	7.60	7.6 8.0; 10.0; 11.3; 13.1; 16.5	10.0
hydraul	Saturated hydraulic condutivity	$m\ s^{-1}$	$1.66x10^{-7}$	$9.99x10^{-6}$	$9.99x10^{-5;}$ $9.99x10^{-5}$; $9.99x10^{-5}$; $9.99x10^{-6}$; $9.99x10^{-7}$; $9.99x10^{-8}$

In general, the simulations reproduced the observed weekly or biweekly variability in the WTD reasonably well, indicating distinct seasonality and interannual variations over the studied years, with primarily very shallow WTDs in the valley environment (Figure 5a). Seasonal and interannual variability are a response to the seasonality of precipitation at the study site (Figure 5b). As soon as the rainy season begins in November, the simulated WTD decreases, with minimum values generally recorded in May, according to observed data. On the other hand, the maximum simulated WTD were recorded at the end of the dry season (September-October) or in the early wet season of next year (November), consistent as observed data.The observed data showed two strong drought events with maximum reduction of the depth of the water table, during the period analyzed. In the early wet season of 2003, the water table fell below 0.61 m from the surface (Table 2). This decrease was due to the lower annual precipitation (33% less than 2002), which was associated with the occurrence of a moderate El Niño event that influenced the Amazonian region between the end of 2002 and

2003 (Climanálise, 2003). Previous studies have documented the association between El Niño events and precipitation reduction in Amazonia (Marengo, 2004). In the dry season of 2003, the mean observed values were underestimated by the model by less than 9%. Another drought event occurred two years later, when a significant rainfall deficit resulted in an exceptional drought during the dry season of 2005 (Marengo *et al.,* 2008; Tomasella et al., 2010), with values up to 50% below the climatological mean in August (Figure 5b).

During this period, the WTD observed fell below 0.54 m from the surface and the model overestimated the observed WTD by approximately 40%. Although WTD showed a larger decrease during the dry season of 2003, the data indicate a higher rainfall deficit in 2005 when the total accumulated was only 393.6 mm, about 32% lower than the presented value during the same period in 2003. This result suggest for this year the INLAND model "dried out" more rapidly than indicated by the observations, highlighting the limited capacity of INLAND to reproduce the abrupt change from wet to dry conditions for the studied area. On the other hand, in the normal years the INLAND simulated very well the transition between wet to dry conditions, indicating a good agreement with observation data. The minimum WTD simulated was found during the wet season of 2004/2005, about 0.08m, while in the same period the observed data indicated the water table closer the surface at 0.01m. The observed and simulated mean WTD in all periods were 0.21 and 0.22 m, respectively, below the surface during the rainy season (November-May), increasing to 0.26 and 0.31 m below the surface at the onset of the dry season (June-October) (Table 2). In general, the observed values were slightly overestimated by the model (Table 3), except in the wet season of 2002/2003 (RMSE = 0.08 m; R^2 = 0.65) and dry season of 2003 (RMSE = 0.06 m; R^2 = 0.79). The overestimation was higher in the dry season of 2005 (0.11 m) with higher RMSE (0.18m) and lower R^2 (0.52). On the other hand, a better agreement was found during the wet season of 2004/2005 with biases of 0.01 m, RMSE of 0.07 m and R^2 of 0.76. The overestimation of the INLAND simulations during 2002-2006 was less compared to that of the Community Land Model (CLM) presented in Fan and Miguez-Macho (2010) at the same study site. According to the authors, the CLM model produced a water table that was 2 m too low compared to the observations. Cuartas et al. (2012) also verified an overestimation of the WTDs simulated using the Distributed Hydrology Soil Vegetation Model (DHSVM) at the same study site, with an RMSE of 0.25 m during both the wet and dry seasons (2002-2006) and an R^2 of 0.60 and 0.72 in the wet and dry seasons, respectively.

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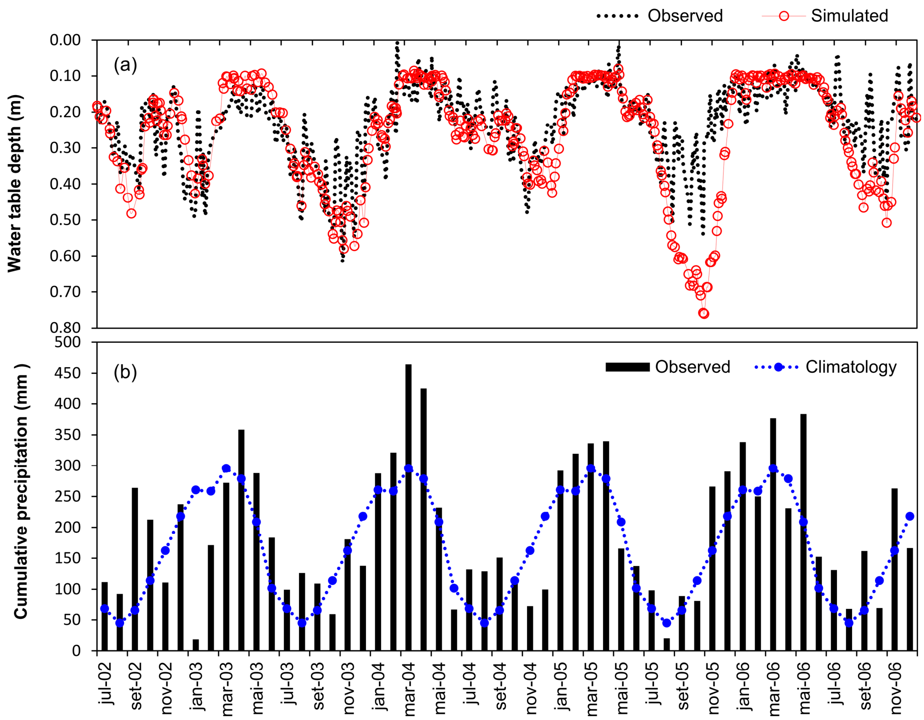

Figure 5. (a) Observed and simulated water table depth fluctuations in the valley area from July 2002 to December 2006**: (b)** Observed precipitation during the same period (vertical bars) and climatological mean precipitation between 1901-1999 in the city of Manaus (dashed line and blue dots).

Table 2. Mean, maximum and minimum water table depth from 2002 to 2006, in the valley area.

	Water table depth (m)					
	Observed			Simulated		
	Mean	**Maximum**	**Minimum**	**Mean**	**Maximum**	**Minimum**
Wet season 2002-2003	0.25	0.49	0.12	0.20	0.43	0.10
Wet season 2003-2004	0.22	0.61	0.01	0.25	0.58	0.09
Wet season 2004-2005	0.20	0.48	0.01	0.21	0.43	0.08
Wet season 2005-2006	0.16	0.54	0.04	0.23	0.76	0.09
Total	**0.21**	**0.61**	**0.01**	**0.22**	**0.76**	**0.08**
Dry season 2002	0.26	0.41	0.16	0.31	0.48	0.18
Dry season 2003	0.33	0.53	0.12	0.33	0.55	0.09
Dry season 2004	0.22	0.34	0.12	0.23	0.30	0.10
Dry season 2005	0.29	0.51	0.14	0.40	0.71	0.10
Dry season 2006	0.22	0.45	0.04	0.29	0.51	0.10
Total	**0.26**	**0.53**	**0.04**	**0.31**	**0.71**	**0.09**

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Table 3. Performance of the INLAND model in representing the depth of the water table in lowland forest, located in the study area, during the period from 2002 to 2006.

	Water table depht (m)		
	Bias	RMSE	R^2
Wet season 2002-2003	-0.05	0.08	0.65
Wet season 2003-2004	0.03	0.09	0.68
Wet season 2004-2005	0.01	0.07	0.71
Wet season 2005-2006	0.07	0.18	0.52
Total	**0.02**	**0.11**	**0.46**
Dry season 2002	0.04	0.07	0.66
Dry season 2003	-0.003	0.06	0.79
Dry season 2004	0.01	0.04	0.57
Dry season 2005	0.11	0.18	0.52
Dry season 2006	0.07	0.10	0.58
Total	**0.05**	**0.11**	**0.54**

b. Soil moisture characteristic curves on the plateau

Both the magnitude and seasonal amplitude variations are apparent in the model simulations and observations throughout the monitored soil profile from December 2001 to December 2006 (Figure 6). The observational data for the deepest layer (obtained by TDR), which was obtained at an hourly temporal frequency, is represented here using weekly or biweekly temporal frequencies to match the observational data from the first 5 layers (neutron probe) from January 2003 to February 2006. Six soil parameters (Table 1) or the model were modified according to available literature in the study region (Tomasella and Hodnett, 1996; Ferreira et al., 2002; Marques, 2009). The sensitivity of the model to these soil parameters was examined through several simulations using different soil parameter values. Based on our simulations, the three most significant parameters were porosity (*poros*), Campbell's parameter (*bex*) and hydraulic conductivity (*hydraul*). For the baseline simulations, the *poros* and *bex* values were allowed to vary with depth according to measurements reported in Table 1. The corresponding statistical performances indices are included in Table 4. The simulated soil moisture content followed an annual cycle, with decreasing amplitude in the deep layers. In the first 4 soil layers, the variability in the soil moisture between the dry and wet season was more pronounced, compared to the other layers. This behavior was related to a rapid response to precipitation in the top soil layers and to high macroporosity. At greater depths, the response to rain events was damped, particularly in the last 2 layers, which responded only to the seasonal variability in precipitation.

In general, the model underestimated the soil moisture content with an RMSE of 0.017 m^3 m^{-3} and R^2 of 0.54 for the first layer and RMSE of 0.006 m^3 m^{-3} and R^2 of 0.71 for the sixth layer. The RMSE decreased and R^2 increased with depth to the sixth layer, suggesting that minor errors occurred between the fifth and sixth soil layers, where the smallest soil moisture variation was observed. This behavior can be explained by the high microporosity and low permeability of the deep soil layers, which are responsible for slow percolation at deeper depths (Nortcliff and Thornes, 1981). The lower variation in soil moisture on a seasonal scale in the last 2 layers may reflect a minor presence of roots at this depth and less water. This finding indicates that there was no significant extraction of water from the roots at these depths because there was sufficient water in the layers above 400 cm during most years (Broedel et al., 2017). The sixth layer, for example (depth of 4.00-8.00 m), had a nearly constant water content of 0.57 m^3 m^{-3}. An interesting behavior that was noticed in our simulations was the underestimation of soil moisture in the dry season of 2005 in all of the layers of the profile, suggesting that the model had difficulty simulating the extremely dry conditions of the study area during this period. The observational data also indicated that the depletion of soil moisture was more pronounced and lasted longer during the 2005 dry season over the entire soil profile because root uptake during that period was more intense than in normal years. Discrepancies in the water content at the end of the dry season of 2005 between the observational and simulated results could be due to differences in the actual versus assumed root distributions.

Table 4. The RMSE, bias and R^2 calculated from weekly or biweekly soil moisture data.

Layer (m)	Nº	RMSE (m^3m^{-3})	Bias (m^3m^{-3})	R^2 -
0.0-0.2	1	0.017	-0.004	0.54
0.2-0.5	2	0.014	-0.001	0.65
0.5-1.0	3	0.012	0.003	0.70
1.0-2.0	4	0.008	-0.002	0.79
2.0-4.0	5	0.007	-0.004	0.72
4.0-8.0	6	0.006	-0.003	0.71

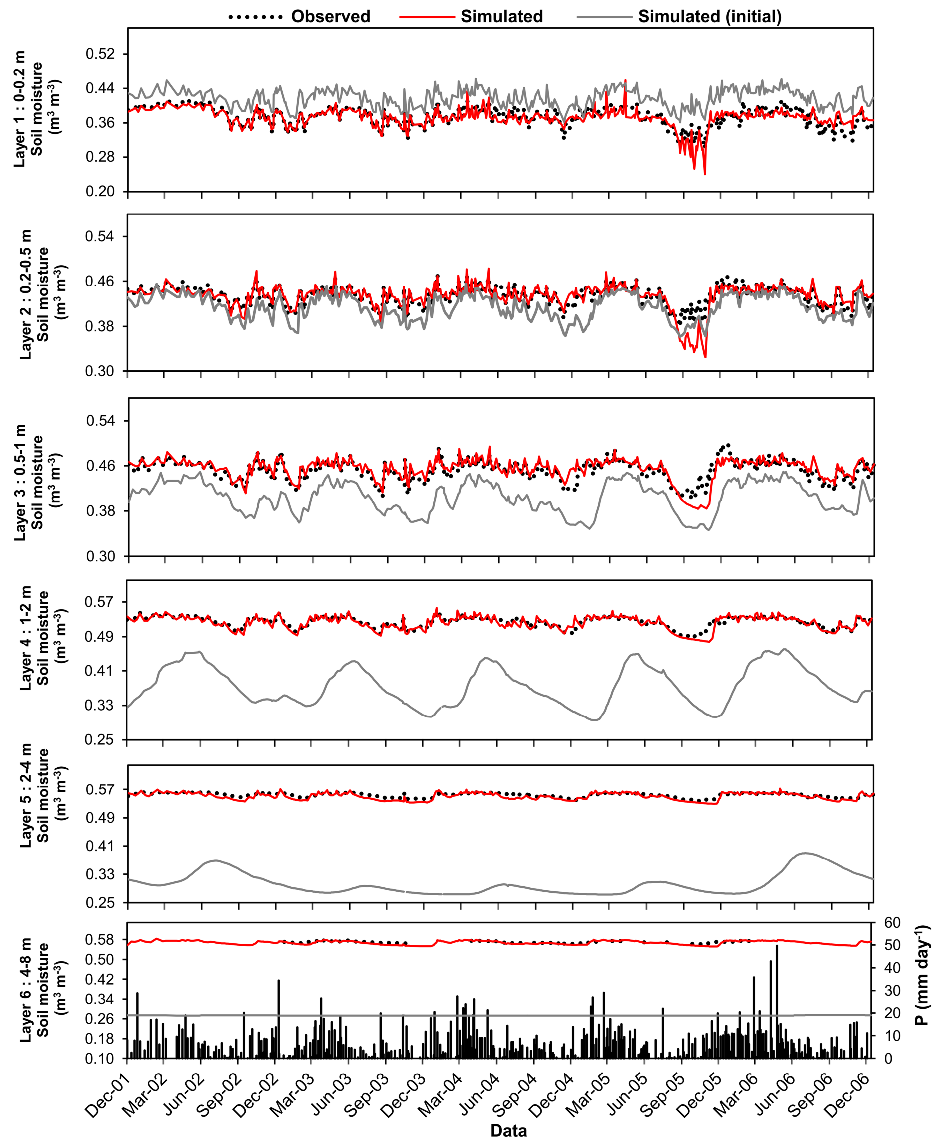

Figure 6. Observed and simulated soil moisture results for the full profile on the plateau. The observational data have a weekly or biweekly temporal frequency. Layers 1- 5: from December 2001 to December 2006, obtained from neutron sonde; Layer 6: from January 2003 to February 2006, obtained from TDR.

c. *Water balance on the plateau and in the valley*

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The INLAND model simulated very well the difference of ET fluxes between plateau and valley, showing larger values on the plateau than that in the valley, in accordance with the observed data in 2006 (Figure 7a). The simulated annual mean ET was 3.7 mm day^{-1} and 3.2 mm day^{-1} for the plateau and the valley, respectively, values close to the observed data, from 3.0 mm day^{-1} in the plateau and 2.9 mm day^{-1} in the valley. Similar results for the plateau area in the same region of study during 2000-2008 were found by Broedel (2012) - 3.6 mm day^{-1} in the calibration of the CLM model - and by Assunção (2011) - 3.5 mm day^{-1}. In addition, the mean ET found in this study for the plateau area also corroborates both the ET value of 3.8 mm mm day^{-1} estimated by Tomasella et al. (2008), for a four years subset of the data series observed, using the Penman-Monteith method, and a value of 3.7 mm day^{-1}, reported by Shuttleworth (1988). The difference in the ET along a topographic profile was also verified by Hodnett et al. (1997a), who estimated ET in two different locations in a forest in the central Amazon, i.e., on a plateau and slope (between a plateau and valley) using the water balance method and found values of 3.8 mm day^{-1} and 3.6 mm day^{-1}, respectively. According to Zanchi (2013), the water extraction in poorly drained valleys occurs at a shallower depth and can be reduced compared with that on plateaus. Hodnett et al. (1997a) determined that the extraction of water from the soil when the water table is near the surface is small (0.5 to 1 mm day^{-1}) compared with the value obtained when the water table is below 1 m depth (3 mm day^{-1}).

The model was also able to simulate the seasonal patterns of the water balance components between the plateau and valley with good accuracy. In both areas, there was a minimum in the simulated ET during the wet season (plateau: 3.3 mm day^{-1}; valley: 3.0 mm day $^{-1}$) that rose in the dry season (plateau: 4.3 mm day^{-1}; valley: 3.6 mm day $^{-1}$), compared to observed data in wet (plateau: 2.5 mm day^{-1}; valley: 2.4 mm day $^{-1}$) and dry (plateau: 3.6 mm day^{-1}; valley: 3.2 mm day^{-1}) season. Da Rocha et al. (2009), in central Amazonia, also described a progressive increase (by 10%) in ET during the dry season when compared to the wet season (2.8 mm day $^{-1}$), and was dominated by a net radiation and vapor density deficit in a plateau area. All the observed ET values were overestimated by INLAND during the entire simulated period, in both areas. However, these estimations are probably lower than the real ET levels because of the problem of the energy balance closure of the eddy covariance systems in tropical forests (Twine et al. 2000; Wilson et al. 2002; Von Randow et al. 2004), mainly in the valley area during the wet season. The bias values indicate a better agreement between simulated and observed curves in the valley area (dry season = 0.4 mm day^{-1}, wet season = 0.2 mm day^{-1})

when compared to the plateau (dry season = 0.7 mm day^{-1}; wet season = 0.9 mm day^{-1}) (Table 5). In the plateau area, the highest R^2 values were found both in the wet season (0.71) and in the dry season (0.64), indicating a better correspondence between observed and simulated data when compared to valley (wet season = 0.63; dry season = 0.56). In addition, RMSE values for the plateau during the dry and wet season were 1.1 mm day^{-1} and 1.0 mm day^{-1}, respectively, values similar to those found in the valley during the dry (0, 8 mm dia-1) and wet (1.1 mm day^{-1}) season.

Mean ET over 2006 corresponded to 55.7% and 45.4% of precipitation in the plateau and shallow areas, respectively. These values are close to those found by Tomasella et al (2008) in the study area of 53%, during the period from 2001 to 2004. In addition, they are also similar to the values of 50% found by Assunção (2011) in the calibration of the IBIS model and 56.9% obtained by Cuartas et al. (2012) using the hydrological model DHSVM, both in the same study area. Of the total loss by ET, 76% were due to transpiration in both forest areas. On the other hand, the evaporation of the intercepted water (interception loss) represented 21.2% and 19.2% in the plateau and valley, respectively, while 2.5% and 3.8% were associated with direct evaporation of the soil, respectively, in the plateau and valley. These results show that in the valley area, water interception by the canopy was slightly lower than in the plateau (10.4%), while soil evaporation was about 52% higher than the plateau. Together these results reflect the structural attributes of the vegetation in this area, characterized by a greater spacing between plants, which shows a more open canopy when compared to the plateau (Nobre et al, 1989), lower trees (Ranzani, 1980; Pinheiro, 2007), a smaller LAI (Cuartas et al., 2012) and different tree composition with less species (Ribeiro et al., 1999; Oliveira and Amaral, 2004). The INLAND model was able to simulate very well the differences between the vegetation in both the areas, including a LAI of 6.1 m^2 m^{-2} in the plateau area and 5.8 m^2 m^{-2} in the basin, in agreement withavailable literature (Cuartas et al., 2012; Marques-Filho et al., 2005). The evaporation of intercepted water corresponded to 11.2% and 9.0% of precipitation, in the plateau and valley area, respectively. These values are comparable to the 11% estimate obtained by Tomasella et al. (2008) for the period 2001-2004, in the same experimental area. The same behavior was also observed in the estimate by Shuttleworth (1988) of 12.4%, carried out in the forest of the Ducke Reserve, in Manaus, near the study area. However, it is higher than the value reported by Lloyd et al. (1988) of 8.9%, in the field experiment also in the Ducke Reserve and the value of 8.2% found by Assunção (2011) in the study area used in the present study. In addition, the values simulated by

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INLAND for plateau and valley are much lower when compared to 16.5% found by Cuartas et al (2007), in the same study area. On the other hand, soil evaporation represented a small part of the total precipitation, only about 1.3% (plateau) and 1.7% (valley), lower than that found by Drucker (2001) of 2.2%, in a hydrological modeling study in the central region of the Amazon, near the experimental site. According to Luizão (1989), there is almost no evaporation of the soil under *terra firme* forest, which explains its lack of consideration in many studies on hydrological modeling in the Amazon, such as that of Tomasella et al. (2008).

The surface runoff (R) simulated by INLAND on both areas was reproduced relatively well by INLAND during the period from 2002 to 2006. Simulated R in the valley was larger over the entire simulation compared with the plateau (Figure 7b). Furthermore, the average simulated daily R rates also indicated low seasonal variability on the plateau compared with the valley. This result is in agreement with the literature, according to which the surface flow rate in the plateau of the study area was low or even absent (Luizão et al, 1989). This is because the higher interception in this area barely reduced the incoming water flux and hence had a minor impact on the R coefficient. The seasonal variability of R rates, which is directly controlled by the seasonal cycle of precipitation, was very well represented by the model, with higher values in the wet season (plateau = 0.4 mm day^{-1}, valley = 2.7 mm day^{-1}) when compared to the dry season (plateau = 0.1 mm day^{-1}, valley = 0.4 mm day^{-1}), in both areas. The INLAND model simulated a peak of R in the valley area during September (0.62 mm day^{-1}), in response to the increase in accumulated mean precipitation in this month (154.9 mm), according to observed data (0.43 mm day^{-1}). This happens due to the proximity of the surface water table to the surface in valley area, which may have induced higher soil moisture, generating saturation, and ultimately leading to overland flow. The mean R observed in the valley area was reproduced relatively well by INLAND with strong overestimated during wet season (bias = 1.7) and slight underestimated in the dry season (bias = -0.1) (Table 5). In addition, during the wet season the RMSE was 2.0 mm dia^{-1} and R^2 of 0.72. In the dry season, a reduction in the value of R^2 (0.40) can be observed, indicating a lower correspondence between simulated and observed R data in the valley while RMSE also exhibited a reduction, dropping to 0.2 mm dia^{-1}. The R represented an average value of 3.9% of total precipitation, during 2002 to 2006, in the plateau area. This value is in agreement with other studies in different regions of the Amazon. The average surface runoff estimated at the Ducke Reserve site in the center of the Amazon by Leopoldo et al. (1995) indicated that only 3% of precipitation was directly lost during storm surges through runoff. Germer et al. (2009) identified a surface runoff of only 1% of the precipitation observed in the southwest of the Amazon, while in the Eastern Amazon, Moraes et al. (2006) reported a coefficient of flow of only 2.7%. On the other hand, in valley area the R simulated by INLAND represented a higher part of precipitation when compared to plateau, about 25.3%.

Deep drainage (D) is also directly linked to precipitation and has a well-defined seasonal cycle, with higher values during the wet season, in both areas, that was very well represented by the INLAND model (Figure 7c). During the wet season, the mean D was 4.3 mm day^{-1} and 2.9 mm day^{-1} in plateau and valley area, respectively reducing to 0.8 mm day^{-1} in the plateau and 0.9 mm day^{-1} in valley area, during the dry season. In addition, the INLAND model simulated a higher D rate in the plateau than in valley area, during the wet season. These differences found between D in both areas occurred because the soil in the plateau is significantly deeper and the water table is located approximately 35 m deep (Tomasella et al., 2008), while in the valley the data showed that the maximum water table depth during the analysis period was approximately 0.6 m. The simulated D in valley area is in accordance with observed data in the same period of 2.4 mm day-1 and 1.6 mm day-1 in wet and dry season, respectively. In general, the INLAND model overestimated R during the wet season (bias 0.5 mm day^{-1}) and underestimated in the dry season (bias = 0.6 mm day^{-1}) (Table 5). The RMSE was 0.7 mm day^{-1} in both areas and R^2 of 0.65 indicating a better correspondence between observed and simulated data during the wet season. D values represent a large portion of the total precipitation, corresponding to a mean percentage of 41.3% and 27.2% in plateau and valley area respectively during 2002-2006. Simulated D values simulated on the plateau are in accordance with the value of 40% found by Broedel (2012), in the same study area during 2000-2008. It is however considerably lower than the value of 29.8% reported by Assunção (2011), in the same area.

Deep drainage added to the runoff results in total runoff (R_{total}). The mean R_{total} simulated during the period from 2002 to 2006 represented 45.2% and 52.5% of the total precipitation, for plateau and shallow area, respectively. The lower percentage of R_{total} in relation to precipitation in the plateau environment is a consequence of a higher ET in this area when compared to the lowland. The mean values of R_{total} in relation to precipitation found in this study, in both areas, showed a good correspondence with the values of 44.3% reported by Tomasella et al. (2008) during the period from 2001 to 2004, and of 49.3% obtained by

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Broedel (2012) between 2000 and 2008 using the CLM model, the same area of study (plateau). The value obtained by Broedel (2012) is identical to the value found by Assunção (2011) in the calibration of the IBIS model. In addition, our results also corroborate with the value obtained by Cuartas et al. (2012) from 2002 to 2004. Using the DHSVM model, Cuartas and colleagues reported that the R_{total} value represented 47.5% of the total precipitation in our study area (plateau).

Table 5. Performance of the INLAND model for the representation of the total evaporation (ET) during 2006; R and D during 2002-2006 located in the study area, from hourly data.

			Wet season			Dry season		
	Flux	**Período**	**RMSE**	**Bias**	**R^2**	**RMSE**	**Bias**	**R^2**
			(mm day^{-1})	(mm day^{-1})	-	(mm day^{-1})	(mm day^{-1})	-
Plateau	ET	2006	1,1	0,9	0,71	1,0	0,7	0,64
Valley	ET	2006	1,1	0,2	0,63	0,8	0,4	0,56
	R	2002-2006	2,0	1,7	0,72	0,2	-0.1	0,40
	D	2002-2006	0,7	0,5	0,65	0,7	-0.6	0,64

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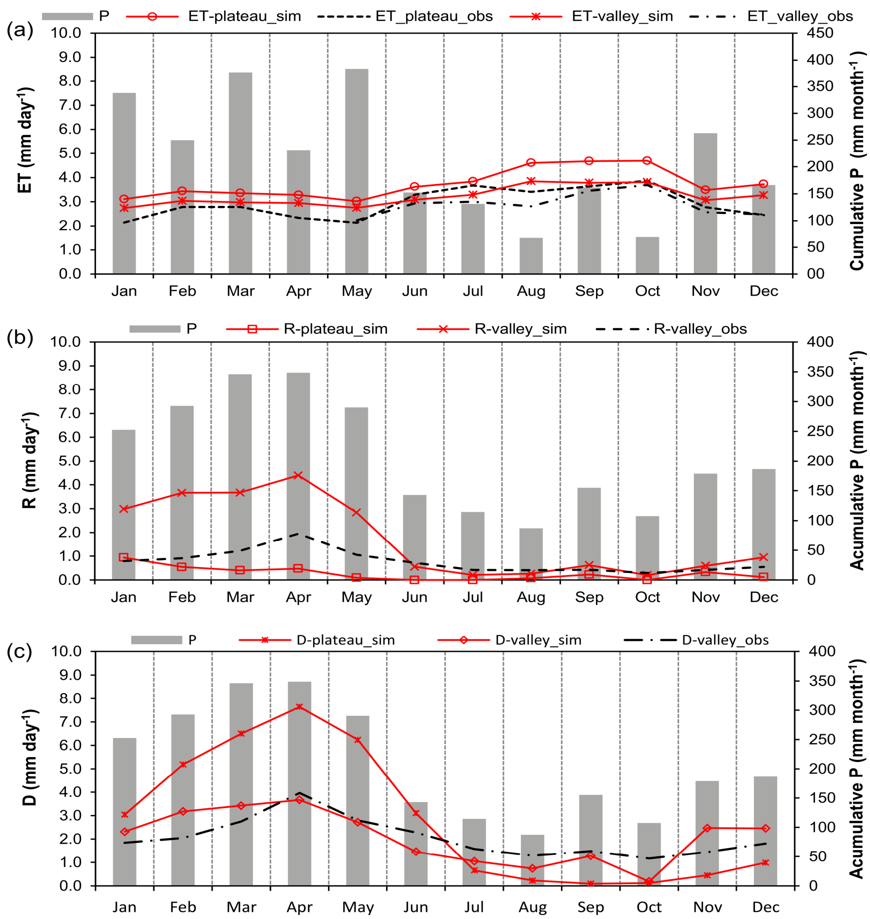

Figure 7. Monthly total observed precipitation (bars), observed ET (continuous lines) and simulated total evaporation, surface runoff and groundwater drainage for the forest canopy (dotted lines).

d. Net Ecosystem Exchange for the plateau and valley

In general, the simulated daytime and nighttime NEE fluxes suggest that the patterns and magnitudes were in agreement with observations on the plateau and in the valley in both the wet and dry seasons. The difference between the two areas was well captured by INLAND using an unconfined aquifer model to simulated the valley area and corrected vegetation and soil parameters in both areas (Figure 8 a, b). During the daytime (dominated by photosynthesis processes), the simulated NEE fluxes on the plateau were slightly higher than in the valley during the wet and dry seasons, in agreement with observed data. The average

NEE flux simulated during the daytime on the plateau, for example, was -8.5 µmol m^{-2} s^{-1} and -10.2 µmol m^{-2} s^{-1} in the wet and dry seasons, respectively, whereas in the valley, it decreased to approximately 11.8% and 23.5% of the values in the plateau, during the wet and dry seasons, respectively (Table 6). In contrast, the observed NEE in the plateau presented a mean of -9.1 µmol m^{-2} s^{-1} and -8.8 µmol m^{-2} s^{-1} in the wet and dry season, respectively, while in valley the NEE fluxes showed a decrease of 25.3% fall in the wet season and 30.7% in the dry season. The observed values are in agreement with Malhi et al. (1998) of 10.2 µmol m^{-2} s^{-1} measured in central Amazonia using eddy covariance, close to study site,. According to Castilho et al. (2006) the higher NEE rate during the daytime on the plateau is suggestive of a stronger CO_2 uptake rate due to more photosynthetic activity resulting from the higher biomass found in this area. According to Laurance et al. (1999), the biomass variation between plateau and valley is mainly associated with nitrogen availability in both areas. The plateau area has a greater amount of nitrogen both in the soil and in the leaves of the plants when compared to the valley, which indicates a greater amount of available resources for the growth of the plants in this area (Luizão et al. 2004). Luizão (1989) conduced a three year long litterfall experiment in an area 10 kilometers to the northwest of study site and showed that on the plateau a 3 year mean of 8.3 tons C ha^{-1} of litter was produced, with only 7,4 tons C ha^{-1} in the valley, indicating that plateau area has higher productivity than the valley. In addition, using sapflow data from Zanchi (2013) for the study site, Stijnman (2015) showed that photosynthesis starts later and ends earlier in the valley area. According to Stijnman (2015), the start and end of sapflow in our study site is linking with solar radiation.

The observed NEE fluxes during daytime did not indicate a pronounced seasonality. In the plateau area for example, the NEE was only 3.3% and 10.2% higher during the wet season in the plateau and valley, respectively. The simulated data show higher mean values of NEE in the dry season, probably due to a higher solar radiation during this period, mainly in the plateau area (20%). Similar behavior was found by Asunção (2011), through the calibration of the IBIS model for the same area of study. The observed daytime NEE values, using eddy covariance, in several areas of Amazonia indicated diverging trends between wet and dry season. The results found by Goulden et al. (2004), for example, in the east of Amazonia indicate higher forest assimilation during the dry season. However, according to Goulden et al. (2004), these results are opposite to those expected, based on forest biomass increment observations, which showed a clear increase during the wet season. In central Amazonia, the data from Malhi et al. (1998) showed higher values in the wet season, related with the rainfall,

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in agreement with observed data from our study site. Despite this, the INLAND model did not represent well the mean seasonality over 2006, but there was a good agreement with the observed daily peak. In the plateau, for example, the maximum values found between 11:00 and 12:00 hours were -20.3 µmol m^{-2} s^{-1} and -16.7 µmol m^{-2} s^{-1} and in the dry and wet season, respectively. These values are close to the observed data in the plateau area, whose peaks occurred between 10:00 and 11:00, -21.9 µmol m^{-2} s^{-1} in the dry season and -21.5 µmol m^{-2} s^{-1} in the wet season. In the valley area, small values of -15.3 µmol m^{-2} s^{-1} and -14.3 µmol m^{-2} s^{-1} were found in the dry and wet season, respectively (between 11: 00-12: 00), values comparable to those observed of -19.9 µmol m^{-2} s^{-1} in the dry season (at 11:00) and -17 µmol m^{-2} s^{-1} in the wet season (at 10:00).

The INLAND model simulated very well the mean NEE fluxes during the night (dominated by respiration processes), showing higher values on the plateau area in both seasons. The mean of 3.2 µmol m^{-2} s^{-1} and 2.9 µmol m^{-2} s^{-1} was found in the wet and dry season, respectively, while in the valley the data indicate a reduction of 18.8% in the wet season and 17.2% in the dry season, values that show good correspondence with observed data of 24% and 13.3% in wet and dry seasons, respectively (Table 6). According to Araújo (2009), there are two possible explanations for this difference. First, due to either dynamically unstable and turbulent flow or neutral atmosphere above the canopy on the plateau (in contrast to dynamically stable and laminar flow above the canopy in the valley), a more efficient vertical transport is favored. Second, the CO_2 stoke fluxes decreased throughout the night in the valley, whereas this was not the case on the plateau. In addition, the lower values of respiration in the fallow, in both seasons, may also be associated with a lower decomposition rate of organic matter in this area. According to Luizão et al. (2004), the canopy of the vegetation in the valley area have lower values of C:N ratio, which confers poor quality to the leaves of the plants and consequently leads to a reduction of the decomposition process. The INLAND model also simulated satisfactorily the seasonality of observed NEE fluxes during the night, which was slightly higher during the wet season, suggesting a higher respiration rate of the plants for the forest during the months with higher precipitation. This increase corresponded to 10.3% and 8.3% in the plateau and valley, respectively, values comparable to observed data, of 20% in the plateau and 5.1% in the valley. In part, this can be explained by increasing of the air humidity that favors an intense organic matter decomposition by the decomposing microorganisms (Luizão and Schubart, 1987). In contrast to the wet season, during the dry season higher temperatures and greater moisture limitation favors the limitation of microbial activity (Malhi et al., 1998), mainly in plateau area. The proximity of the water table depth to the surface in the valley area ensures a humid environment for a good part of the year, even during the dry season. In addition, during the wet season a greater cloudiness and amount of water vapor is observed in the atmosphere, contributing to a lower loss of longwave radiation (Lin) and less cooling of the vegetation canopy. This influences the stability of the atmosphere, which becomes dynamically unstable and consequently favors vertical transport and mixing of CO_2 fluxes by the forest, both in the plateau and valley areas, during the wet season (Araújo, 2009; Miller et al., 2004). Other studies also found higher respiration values during the wet season. Zanchi et al. (2014) found higher soil respiration or Rsoil (microbial and fungal together with plant roots) during the wet season in the study site, for both plateau and valley area. Higher values of Rsoil also was found by Souza (2004) in the plateau area, in the same study site.

In general, all observed NEE fluxes were underestimated by model, mainly during the dry season in the valley area (Table 6). This underestimation, also found by Assunção (2011) in the same study area, is due to the overestimated NEE value obtained by the eddy covariance method, observed since the first tests performed using this methodology (Ontaki, 1984). The eddy covariance method underestimates the flux of CO_2 under stable atmospheric conditions during the night (Miller et al., 2004), when CO_2 flow is characterized by autotrophic respiration and the decomposition of organic material from soil. The drainage of CO_2 from the highest areas, as is the case of the site of tower K34, to the lower parts, located in the bottom of the valley (Tóta et al. 2008; Araújo et al. 2010) also contributes to the underestimation of forest CO_2 flow at night. In addition, the lowest values of R^2 also were founding in the valley area (Table 6), suggesting a lower agreement between the observed and simulated NEE values, especially during the dry season (0.55) when compared to the wet season (0.63). The R^2 values in the plateau (wet = 0.76; dry = 0.66) are higher than those found by Assunção (2011) and Imbuzeiro (2005) of 0.54 and 0.41, respectively, both in the same study area. On the other hand, the RMSE values of the two areas were very similar. In the plateau, for example, it was 6.6 µmol m^{-2} s^{-1} and 7.2 µmol m^{-2} s^{-1} in the wet and dry season, respectively, while in the valley, corresponding values were 6.4 in the season and 7.5 in the dry season. These values are close in magnitude to that reported by Imbuzeiro (2005), of 4.2 µmol m^{-2} s^{-1}, in the plateau area.

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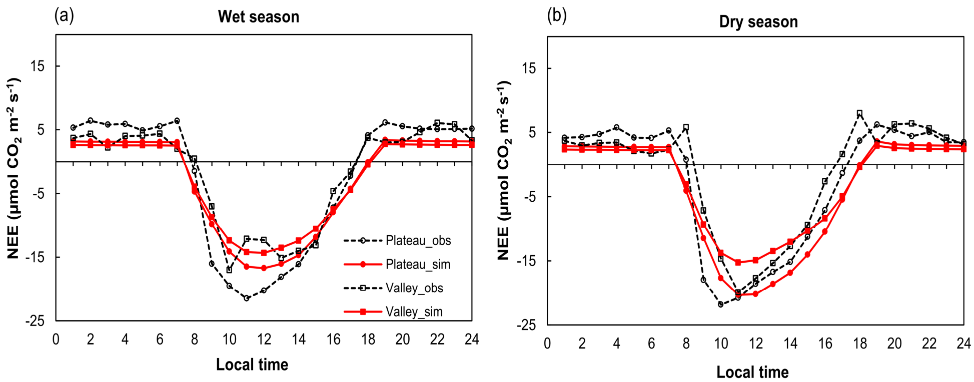

Figure 8. Average daily observed and modeled net ecosystem exchange (NEE) rates between the ecosystem and the atmosphere). During the daytime, negative values represent the uptake of CO2 from the atmosphere (photosynthetic activity is higher than respiration), whereas a positive flux at nighttime is associated with emissions of CO2 from the forest to the atmosphere (respiration activity only).

Table 6. Average observed and simulated net ecosystem exchange (NEE) values for the plateau and valley during the day (06:00–18:00 h), night (18:00–06:00 h) and daily totals. The Daily peak observed and simulated as also includes.

	Wet season				Dry season			
	$\mu mol\ CO_2\ m^{-2}\ s^{-1}$				$\mu mol\ CO_2\ m^{-2}\ s^{-1}$			
	Daytime	Nighttime	Daytime peak	Daily total	Daytime	Nighttime	Daytime peak	Daily total
Plateau-Obs	-9.1	5.4	-21.5	-2.4	-8.8	4.5	-21.9	-2.8
Plateau-Sim	-8.5	3.2	-16.7	-3.2	-10.2	2.9	-20.3	-4.2
Valley-Obs	-6.8	4.1	-15.1	-1.8	-6.1	3.9	-19.9	-1.5
Valley-Sim	-7.5	2.6	-14.3	-2.9	-7.8	2.4	-15.3	-3.1

Table 7. Performance of the INLAND model for the representation of the net ecosystem exchange (NEE) during 2006, from hourly data.

	Plateau		Valley	
	Wet season	Dry season	Wet season	Dry season
RMSE ($\mu mol\ CO2\ m^{-2}\ s^{-1}$)	6.6	7.2	6.4	7.5
Bias ($\mu mol\ CO2\ m^{-2}\ s^{-1}$)	-0.7	-1.5	-0.8	-1.7
R^2 (dimensionless)	0.76	0.66	0.63	0.55

e. Energy fluxes for the plateau and valley

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The simulated Rn diurnal evolution was very similar over the plateau and valley during the wet and dry seasons, which was mainly due to incoming solar radiation and because both have the same meteorological characteristics (Figure 10 a, b). The R^2 of the plateau was 0.99 in both the wet (RMSE of 11.7 W m^{-2}) and dry (RMSE of 11.3 W m^{-2}) seasons, whereas for the valley, it was 0.99 in the dry season (RMSE of 29.8 W m^{-2}) and 0.97 (RMSE 44.7 W m^{-2}) in the wet season (Table 8). In general, the simulated Rn was slightly underestimated in the valley. One possible explanation for this difference is that the simulated albedo in both the wet and dry seasons was higher than the observed values. The constant albedo simulated by INLAND in the valley were 10.5% and 10.8% in the wet and dry seasons, respectively. These values are 5% (wet season) and 3% (dry season) higher than previous observational values (Araújo, 2008; Oliveira, 2010). Furthermore, this difference could be due to reduced incoming longwave radiation, which is derived from meteorological data input in the model in the plateau. According to Araújo (2008), observed incoming longwave radiation is consistently higher in the valley than on the plateau. The INLAND model simulated very well the difference of the albedos in both areas, showing higher values on the plateau (wet season = 12%; dry season = 12.3%) in agreement with observed data (wet season = 11%; dry season = 12%) from Araújo (2008) and Oliveira (2010). The correct adjustment of albedo in both areas was was made through reflectance of near infrared radiation from leaves, i.e., *rhoeveg_NIR* parameter of the 0.31 and 0.26 to plateau and valley respectively (Table 1). According to Leitão (1994), this difference between both areas is caused by the vegetation canopy structure. In the valley area, the foliage of the vegetation grouped in the canopy exhibits larger peaks and depressions organized on the canopy surfaces, and a large amount of incident solar radiation penetrates the canopy before being reflected (Shuttleworth, 1989). In contrast, the vegetation in plateau area is denser which favors a greater homogeneity of the vegetation, resulting in a greater reflectivity of the solar radiation. The seasonality of albedo also was very well representing by INLAND, with higher values during dry season. According to Malhi et al. (1998) the canopy of the forest is darker during the rainy season, due to the seasonality of the precipitation, favoring greater absorption of radiation and consequently lower albedo.

The model was able to accurately simulate the partitioning of available energy (Rn) with higher percentage of Rn destined for LE flux, followed by H flux, also verified by Araújo et al.(2002), Malhi et al.(2002) and Randow et al.(2004).This was possible especially using the vegetation and soil parameters modified for each area (Table 1) and water table dynamic in the valley area. For the plateau, the LE and H annual averages corresponded to 84.9%, 14.8% of Rn, respectively while the corresponding values in the valley were 72.9% and 26.5%. These values were found to agree well with the observations, of 60.4% (LE) and 21.3% (H) for the plateau and 50% (LE) and 21.9% (H) for the valley. The simulated LE was higher on the plateau than in the valley, particularly in the dry season (Figure 10 c, d) according to observed data. One possible explanation for this difference is the higher LAI on the plateau, which produced a higher evapotranspiration rate. According to Marques-Filho et al. (2005), the LAI on the plateau is 6.1 m^2 m^{-2}, whereas in the valley, the value may be approximately 13% smaller (Zanchi et al., 2014). Another possible reason is the excess water in the valley soils, which can produce oxygen-deficient roots and affect plant survival and functioning (Pezeshki and DeLaune, 2012). The simulated LAI for the mentioned period was approximately 6.1 m^2 m^{-2} on the plateau and 5.8 m^2 m^{-2} in the valley. These values were found to agree well with Cuartas et al (2012) for the same study area during the period from 2002 to 2006. The INLAND model overestimated the LE flux on the plateau and in the valley. The overestimation was stronger on the plateau, where we found an RMSE of 63.8 W m^{-2} and R^2 of 0.85 in the wet season and an RMSE of 66.8 W m^{-2} and R^2 of 0.88 in the dry season. In the valley, the RMSE and R^2 were 63.6 W m^{-2} and 0.72 in the wet season, respectively, whereas in the dry season, the RMSE were slightly smaller, i.e., approximately 59.6 W m^{-2} with a R^2 of 0.81, respectively (Table 8). The most likely reason for the overestimation of the simulated LE is the large hourly errors caused by underestimation of the observational LE, which led to the unclosed energy balance. The INLAND model also simulated very well the seasonality of LE, showing higher values during dry season in both areas (plateau = 124.6 W m^{-2}; valley = 103.6 W m^{-2}) when compared to wet season (plateau = 97.2 W m^{-2}; valley = 86.3 W m^{-2}) which can also be clearly visualized in the observed data.

In contrast to LE fluxes simulated by INLAND, the simulated H fluxes are higher in the valley area when compared to the plateau, in both wet (valley = 29.8 W m^{-2}; plateau = 18.6 W m^{-2}) and dry (valley = 39.8 W m^{-2}; plateau = 19.5 W m^{-2}) seasons (Figure 9 e, f). This behavior is in agreement with the observed H fluxes mainly during dry season (valley = 36.7 W m^{-2}; plateau = 33.6 W m^{-2}), indicating a good performance of INLAND. The diurnal cycle also reveals that the seasonality of the H flux was very well reproduced by the model, presenting higher values during the dry season in both areas, in perfect agreement with the observed data. This strong seasonality in H data occurs because, during the dry season, a higher value of incident solar radiation (Sin) is recorded in Amazonia, due to the lower cloudiness in the region (Araújo, 2009). The mean values of incident radiation found by Araújo (2009) in the valley area, were 44.6% higher in the dry season of 2006. The higher values of H in the valley, compared to the plateau, suggest that more Rn was used for the H flux in the valley area during the dry months (consequently less energy for the LE flow).This finding may be due to the smaller aboveground live biomass in the central Amazon forests that grows in bottomland areas such as valleys (Laurence et al., 1999; Castilho et al. 2006). This scenario appears to be the case in our study area, where bigger trees tend to occur more frequently on high and flat areas that are dominated by clay soils (Ranzani, 1980; Oliveira and Amaral, 2004), whereas they are sparser in the bottom of the valley (Nobre, 1989). Figure 9 (e, f) also shows the significant bias in the simulated values of H for the valley; the model overestimated the observed hourly totals by 11% and 2% during the dry (R^2 = 0.78; RMSE = 34.9 W m^{-2}) and wet (R^2 = 0.66; RMSE = 34.3 W m^{-2}) seasons, respectively. This overestimation in the H flux was related to one of the most relevant parameters for the simulation of H, *rhoeveg_NIR* (Varejão et al. 2011). The sensible heat flux from vegetation to the air is a function of the leaf and stem temperatures and these two variables are dependent on the solar radiation absorbed by the canopies and soil as well as on the net absorbed fluxes of infrared radiation.

Table 8. RMSE, bias and R^2 calculated for the hourly energy and NEE fluxes.

		Wet season			Dry season		
		RMSE (W m^{-2})	Bias (W m^{-2})	R^2 -	RMSE (W m^{-2})	Bias (W m^{-2})	R^2 -
Plateau	Rn	11.7	-4.3	0.99	11.3	-2.2	0.99
	LE	63.8	-34.0	0.85	66.8	-32.6	0.88
	H	26.4	2.8	0.75	33.4	14.2	0.78
Valley	Rn	44.7	15.0	0.97	29.8	10.8	0.99
	LE	63.6	-16.5	0.72	59.6	-20.8	0.81
	H	34.3	-11.0	0.66	34.9	-2.0	0.78

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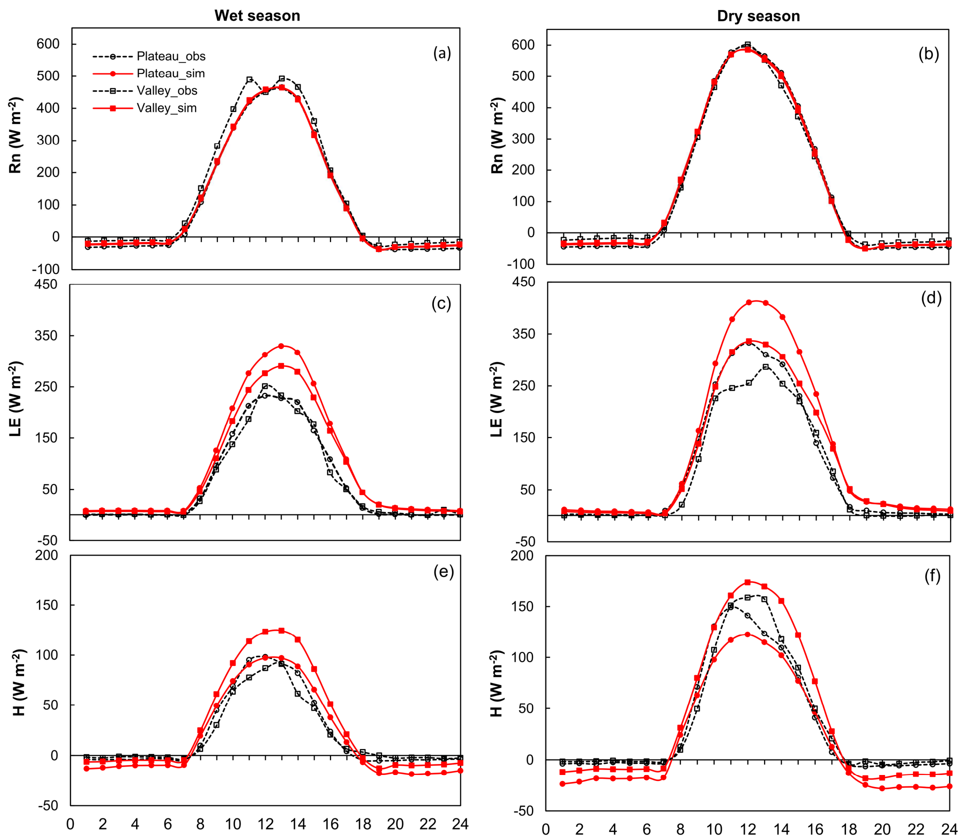

Figure 9. Observed and modeled energy fluxes on an hourly basis in the wet and dry seasons of 2006.

g. Biomass and productivity of vegetation on the plateau and in the valley

The results of the analysis suggest that the INLAND model reasonably reproduced the biomass (leaves, wood and root) and carbon flux variability over the plateau and valley (Table 9). The biomass stocks and carbon fluxes on the plateau were greater than in the valley over the 100 years of simulation. The aboveground biomass stock (leaves and wood) on the plateau in the *DV1* simulation was 172.6 tons C ha^{-1}, whereas in the valley, it was approximately 124.3 ton C ha^{-1}. The aboveground biomass stocks in the second simulation set, *DV 2*, showed a reduction of 28% on the plateau and 38% in the valley. This finding is not surprising because the *DV1* simulation had more initial biomass than *DV2.* The aboveground biomass simulated by INLAND on the plateau and in the valley was similar to the value obtained by Higuchi et al. (2016), i.e., 188 tons C ha^{-1} in an area of 3 ha of forest located near our site. The value identified by Malhi et al. (2006) in the same region (10 ha) was 148 tons C ha^{-1}.

According to Malhi et al. (2006), the aboveground biomass in the eastern Amazon is equivalent to 150-175 tons C ha^{-1}. Our simulations agree with the stock value of 167 tons C ha^{-1} estimated by Pyle et al. (2008) for the central Amazon for 20 ha of forest, same value of estimative from Johnson (2016) to central Amazonia. In all of these studies, the aboveground biomass was considered as the mean of the entire transect that encompasses a plateau-valley mosaic. The differences in aboveground biomass between the plateau and valley is in agreement with several studies in the region (Castilho et al. 2006; Zarin et. al 2001, Laurence et al 2001; Luizão et al. 2004) indicating good performance of INLAND. Some of these differences are due to the forest structure, as plateau areas have a greater proportion of large trees than valley bottoms (Malhi et al., 2009b). Furthermore, the biomass is influenced by soils, and more fertile soils tend to favor fast-growing plants and low wood density (Malhi et al. 2004), which allocates more energy to wood and leaf production and less to structural and chemical defenses and their associated metabolic costs (Malhi et al. 2009b). Castilho et al. (2006), for example, found that tree biomass tended to increase in clay-rich soils (located in plateaus), whereas sandier soils (located in valleys) are characterized by lower productivity (Zarin et al. 2001). Laurance et al. (1999) attributed the great spatial variation on aboveground biomass estimates to nitrogen availability. The clayey soils located on high plateau areas are considered more fertile and are characterized by higher nitrogen availability and decomposition rates (Luizão et al. 1989), which provide greater productivity to the forest, indicating that higher carbon fluxes can be found on plateaus and decline toward valley bottoms.

Coarse and fine roots of live trees compose the belowground biomass stocks. However, the belowground biomass simulated by INLAND is defined only by the fine rootstock, which includes root material less than a threshold diameter, usually 2 mm. According to Malhi et al. (2009 b) the fine roots are a very minor component of the belowground biomass stock, about 9.5% of aboveground biomass. The simulated values for both *DV1* and *DV2* were higher on the plateau compared with the valley. The fine root biomass simulated in *DV1* was 1.9 tons C ha^{-1} and 1.6 ton C ha^{-1} on the plateau and in the valley, respectively. The *DV1* simulation represented a large reduction of 26% on the plateau and even stronger in the valley, approximately 37%. In general, the model correctly simulated the variability of the fine root biomass over the plateau and valley, with higher values on the plateau. Root biomass would be expected to be low in soils limited by anoxia associated with seasonally higher water tables, such as those found in the valley (Malhi et al., 2009b). However, the simulated results in both environments were underestimated when compared with those of other studies in the central and eastern Amazon, which were realized along a topographic profile in 1 meter of soil. Malhi et al. (2009b) estimated a value of 42 tons C ha^{-1} belowground biomass (coarse and fine roots) in a forest in the central Amazon, whereas in the east, this value ranged from 35 to 48 tons C ha^{-1}. In all these studies, the fine roots showed values of 4 tons C ha^{-1}. A study by Metcalfe et al. (2007) in the same area to the east of the basin reported much lower values of approximately 2 tons C ha^{-1}. The values of belowground biomass in similar forests located within 100 km of one another can exhibit significant differences, and it is not clear what results in these differences (Chambers et al. 2001b).

The INLAND model simulates both autotrophic (Ra) and heterotrophic (Rh) respiration terms that together represent the total ecosystem respiration (Reco). The simulated values of Rh (CO_2 respired by herbivores, detritivores, and higher trophic levels as they consume and break down organic matter) and Ra (CO_2 directly respired by plants - leaf, wood and root live- as a breakdown product from their own metabolic activity) rates were not significantly different between the *DV1* and *DV2* simulations. The Ra flux for *DV1* simulations varied from approximately 24.9 tons C ha^{-1} yr^{-1} to 20.1 tons C ha^{-1} yr^{-1} on the plateau and in the valley, respectively. These results are similar to the estimates reported by Malhi et al. (2009b) and Chambers et al. (2004) in the same region (mean of the entire transect) of approximately 19.8 tons C ha^{-1} yr^{-1} and 21 tons C ha^{-1} yr^{-1}, respectively. These values also are close to estimates of 21.4 tons C ha^{-1} yr^{-1} by Malhi et al (2009) in eastern Amazonia. Joetzjer et al. (2015) found higher values, of 25-32 tons C ha^{-1} yr^{-1} in central Amazonia using the land surface model ISBACC (Interaction Soil Biosphere Atmosphere Carbon Cycle), from Noilhan and Mahfouf (1996). Rh was significantly lower than Ra, with values for *VD1* simulations varying from 9.4 tons C ha^{-1} yr^{-1} on the plateau to 9.1 tons C ha^{-1} yr^{-1} in the valley. Estimates of Rh from this study were comparable to those previously reported by Malhi et al. (2009b) of 9.6 tons C ha^{-1} yr^{-1} and Chambers et al. (2004) of 8.5 tons C ha^{-1} yr^{-1}.

The Rh (microbial and fungal) together with Ra (plant roots) composed efflux of carbon dioxide (CO_2) from soil or soil respiration (Rsoil). According to Chambers et al. (2004) there is a strong correlation between Rsoil with soil texture, which varied along the topography. The root respiration, for example, is higher in the plateau (7.4 tons C ha^{-1} yr^{-1}) when compared to the valley (5.8 tons C ha^{-1} yr^{-1}) area (Matcalfe et al. 2007). This result is opposite that found by Silver et al. (2005), also in eastern Amazonia (plateau = 3.2 tons C ha^{-1} yr^{-1};

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valley = 5.2 tons C ha^{-1} yr^{-1}), using a very different methodology. The ratio of root respiration to total Rsoil may vary from 50 to 60% depending on vegetation type and season (Hanson et al., 2002). Chambers et al. (2004) estimated that root respiration accounted for 45% of Rsoil in our study area. This value is higher that found by Silver et al. (2005) in east of Amazon, of 24-35%. The values of total Rsoil reported by Chambers et al (2004) were 14.4 tons C ha^{-1} yr^{-1} for plateau (clayey soils) and 9.8 tons C ha^{-1} yr^{-1} for valley (sandy soils), suggesting a good performance of the INLAND model in representing Rh. This result can be explained by lower decomposition rates in valley locations (Luizão et al, 2004). According to Luizão et al, (2004), besides the annual production of litter in valley (6.6 tons C ha^{-1} yr^{-1}) is lower than in plateau (8.9 tons C ha^{-1} yr^{-1}), similar to result found by Luizão and Schubart (1987), the higher C:N ratio in the leaves of plants in valleys indicate low quality, suggesting that their decomposition rates may be lower than on plateaus (also reported by Souza, 2004). In addition, it is clear from the early studies, that higher water contents can inhibit CO_2 production in soils (Linn and Doran, 1984; Oberbauer et al., 1992; Sotta et al., 2004; Davidson and Janssens, 2006). Sotta et al. (2004), for example, verified a steep 30% drop of Rsoil after rainfall when compared with nonrain periods. During the year of 2006 it is possible to notice that the precipitation in the study area was higher, in almost all months, than the climatological mean in Manaus (Figure5). The total accumulated precipitation during this period was 2591 mm, about 20% higher than the climatological mean. Because of the higher precipitation in valley areas, the high phreatic level and exfiltration of groundwater provoked slow diffusion of oxygen into the soil, presumably only allowing for anaerobic decomposition with generally slower degradative enzymatic pathways. The results of Rsoil from this work are different from those of Zanchi et al (2014) that indicate higher rates on the valley (15.5 tons C ha^{-1} yr^{-1}) when compared to the plateau (9.8 tons C ha^{-1} yr^{-1}) area. However, during the period of Rsoil measurements of Zanchi and co-authors from a plateau area (period completely different than Rsoil measurements in the valley area), which were undertaken between August 3 until November 6 (2006) and February 21 until February 26 (2008), the precipitation was 45.6% and 50.4% higher than the climatological mean in Manaus (Figure 5) indicating a higher replacement of the air filled pores by water that may form a cap and prevent gas diffusion of CO2 through the soil to the atmosphere.

One interesting feature to note is the link in INLAND between Rh and NEE or an exchange of carbon as gaseous CO_2 between ecosystems and the atmosphere. The lower value of Rh in the valley can also produce a smaller simulated NEE in this environment. The NEE simulated by

INLAND ranged from -0.6 to -0.7 tons C ha^{-1} yr^{-1} (for *DV2* and *DV1*, respectively) in the valley, i.e., approximately 40% less than that simulated on the plateau, indicating a lower net CO_2 flux in the forest canopy in the valley forest. However, the NEE simulated by INLAND in both environments are generally very similar to the values found by Malhi et al. (1999) of -1.1 tons C ha^{-1} yr^{-1}. The NEE was also influenced by changes in the NPP over time, by sequestering atmospheric CO_2, and supplies organic material for Rh. The NPP values simulated in the present study on the plateau, i.e., approximately 10.6 tons C ha^{-1} yr^{-1} (*DV1* and *DV2*), were slightly larger than in the valley. The NPP reductions in the valley represented 7% and 8% for *DV1* and *DV2*, respectively, and were similar to the estimate provided by Malhi et al. (2009b) for the central Amazon using flux tower data (10.1 tons C ha^{-1} yr^{-1}). Furthermore, the simulated NPP was also comparable to the observed average of 9.0 tons C ha^{-1} yr^{-1} from a forest in the same study region (Malhi et al. 2009b). According to Aragão et al. (2009), the total NPP tends to increase with soil phosphorus (P) and leaf nitrogen (N) status. It is well known from Ferraz et al. (1998) and Luizão et al (2004) that both soil P and leaf N is higher in plateau areas, where the soil has higher clay content. Simulating the NPP together with the GPP in forest biomes is fundamental for realistic global and regional carbon budgets and for projecting how these fluxes are affected by a changing climate (Zha et al. 2013). As expected, the GPP in the plateau was higher than in the valley and area, without significant difference between both *VD1* and *VD2*. The GPPs on the plateau and valley were 36 tons C ha^{-1} yr^{-1} and 30 tons C ha^{-1} yr^{-1}, respectively, agreeing with the estimate of 30.4 tons C ha^{-1} yr^{-1} provided by Malhi et al. (1998) close to study site. Similar values were also reported by Malhi et al (2009b) for the same study area, of 29.9 tons C ha^{-1} yr^{-1}, and by Fischer et al (2007) in a forest located at east of Amazonia, with a mean value of 31.2 tons C ha^{-1} yr^{-1}. All the references given here, included both topographical classes (plateau and valley). This reduction of 16.7% in the simulated GPP in the valley can be related with the difference of the LAI and ET in both areas. Since the valley LAI values are lower than those in the plateau and consequently so is ET, this should reduce the GPP and produce a large difference between the plateau and valley area. In addition, in the valley area the sutured condition of this environment could lead to a lower GPP, due to the impact of this condition in the process of transpiration (Feddes et al.,1978), resulting in lower productivity of the forest in this area.

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Table 9. Average of the last 10 years of the simulations for the carbon fluxes and biomass on the plateau and in the valley of: GPP (gross primary productivity), NPP (net primary production), NEE (net ecosystem exchange), Rh (heterotrophic respiration), Ra (autotrophic respiration), leaf biomass, wood biomass and root biomass. The *VD1* (dynamic vegetation 1) and *VD2* (dynamic vegetation 2) values correspond to the dynamic vegetation and dynamic vegetation-cold start simulations, respectively.

Variable	Unit	Plateau		Valley	
		VD1	VD2	VD1	VD2
GPP	ton ha^{-1} yr^{-1}	35.5	35.4	29.9	29.9
NPP	ton ha^{-1} yr^{-1}	10.6	10.6	9.9	9.8
NEE	ton ha^{-1} yr^{-1}	-1.2	-1.0	-0.7	-0.6
Ra	ton ha^{-1} yr^{-1}	24.9	24.8	20.1	20.0
Rh	ton ha^{-1} yr^{-1}	9.4	9.5	9.1	9.2
Leaf biomass	ton ha^{-1}	2.9	2.2	2.4	1.5
Wood biomass	ton ha^{-1}	169.7	122.5	121.9	75.8
Root fine biomass	ton ha^{-1}	1.9	1.4	1.6	1.0

4. Conclusions

In this study, we demonstrate that the INLAND model is able to reproduce the particularities of the observed energy, water and carbon fluxes with reasonable accuracy between two very different environments, plateau and valley bottom, in a central Amazonian forest, confirming our initial hypothesis. To better represent their characteristics, an adjustment of vegetation and soil parameters and the development of a lumped unconfined aquifer model was required to account for the water table dynamics in the valley area. The model reasonably reproduces (R^2=0.69) the observed temporal variability of the water table depth in the valley, including its seasonal cycle and interannual differences, in accordance with precipitation variability. On the plateau, the mean ET was 13% higher than in the valley, with maximum amplitude in August, when the precipitation is lower. The surface runoff is significantly higher in the valley bottom during the wet season. The mean surface runoff on the plateau represents a small portion (3%) of the total precipitation over the entire period. In the valley, this percentage is much larger, representing approximately 25% of total precipitation.

The INLAND model also captured the difference in the partitioning of incoming energy and carbon fluxes, mainly LE, H and NEE fluxes in both seasonal and diurnal time scales. On the plateau, the simulated LE flux is higher than in the valley during both the wet and dry seasons, due to its higher biomass stock, allowing more energy to be used for the LE flux. The opposite behavior is observed for the simulated H flux, as in the valley bottom area lower

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stock biomass favor less energy used in LE flux and consequently a higher H flux in both seasons. A large difference between the wet and dry seasons is observed in our results, with much higher H values during the dry season, when more radiation is incident to the surface. The seasonal variability also shows that the values of LE are higher during the dry season than in the wet months, in both areas. In general, NEE is higher on the plateau during both the wet and dry seasons. During the dry season, when simulated NEE is higher, the reduction in the valley is 26%, with a value of only 9.2% during the wet season. Our results suggest that the model also reasonably reproduces the biomass and carbon flux variability between the plateau and valley. The biomass values show significant differences between the two environments. In general, the dynamic vegetation module using the aboveground biomass pool (leaves and wood) is much smaller in the valley than on the plateau after 100 years of model simulation. The biomass reductions in the valley represent 28% and 38% of the two sets of simulations, i.e., *DV1* and *DV2*, respectively. The simulation of belowground biomass does not reproduce well the values observed in other studies in the region. However, all of the simulated fluxes are higher on the plateau than in the valley, suggesting the larger productivity of the plateau forest, in agreement with the available studies in the central and eastern Amazon.

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